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EXPERIMENTS TESTING SEISMOGRAPHIC METHODS FOR DETERMINING CRUSTAL STRUCTURE

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Introduction

This paper sets forth the results of experiments executed with a view to ascertaining the effectiveness of seismographic methods in the determination of the structure of both the more superficial and the deeper parts of the earth's crust.

The methods employed are similar in principle to those utilized for several decades in the investigation of earthquakes. Oscillations—miniature earthquakes—are generated in the crust by detonating buried explosives and are recorded at suitable distances by portable seismographs. The records or seismograms indicate with an accuracy of approximately 0.001 second the relative times of the explosion and of the arrival of the resulting waves.

The arrival times which are recorded most clearly are those of the direct compressional waves, the surface waves, the compressional waves reflected from some depth by boundary surfaces between rock bodies possessing different physical constants, and the compressional waves which, leaving the explosion point with a somewhat steeper path, are refracted through deeper layers having higher velocities and emerge at the recording instruments at a somewhat steeper angle. With considerable distance between explosion and recording points, the latter waves often arrive first in spite of their longer path, due to the high velocities in the deeper formations. Transverse and surface waves are not extensively utilized in these investigations.

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Inferences regarding subsurface structures are based mainly on data derived from the reflected and the refracted waves. Since both the field procedure and the calculations in respect to these two sets of waves differ somewhat, they are commonly referred to as the reflection and the refraction methods.

These two seismic methods have of course been used by oil companies for a number of years in their search for productive areas, but the types of structures to which they have been applied have been limited somewhat by the economic interests of the petroleum geologists. Doubtless because of a disinclination to make the exact nature of the methods and instruments a matter of common knowledge, comparatively little has been written about them, and very few papers have dealt with the effectiveness of the methods for determining crustal structure.

The general principles involved in the design of the instruments, procedure in registration, and interpretation of seismograms are quite similar to those utilized in the study of earthquakes. Portability and ruggedness are of course emphasized in the instruments, and only relative and not absolute time is required.

The investigations discussed in this paper resulted from plans formulated by Dr. Arthur L. Day, director of the Geophysical Laboratory of the Carnegie Institution of Washington. About three years ago Dr. Day, after hearing reports regarding the effectiveness of seismic methods for discovering oil structures, as related at a meeting of the American Association of Petroleum Geologists, indicated that it would be very desirable to test these methods on a wide range of crustal structures and that if they are found to be generally effective for determining the architecture of the crust they would constitute a very powerful new tool for the structural geologist. An opportunity to make this test came in the early summer of 1931, when Geophysical Service, Inc., of Dallas, Texas, through the generous attitude of Dr. J. C. Karcher, offered trained personnel and a complete set of instruments. A grant from the Carnegie Institution of Washington, made available through the broad interest of President John C. Merriam in seismologic research, permitted the prosecution of the investigation.

The field party consisted of Mr. Henry Salvatori, geophysicist in charge; Frank Bierend, instrument operator; Edward Partain, explosives technician; Ray Felton, permit negotiator; and one to three laborers. The authors devised the general plans for the field experiments and participated for considerable periods in the field operations. The transportation of personnel, instruments, explosives, and other equipment involved the use of one truck and two to three automobiles. The field work was done in the months of July and August, 1931.

Some of the problems on which it was felt desirable to test the seismic methods were: (*a*) velocity of earthquake waves in different types of rock; (*b*) depths to successive rock boundaries and thicknesses of rock units calculated therefrom; (*c*) determination, under a considerable area, of the form of the upper surface of a formational unit in a folded district, by both the reflection and the refraction methods, and the structure inferable therefrom; (*d*) determination of the thickness of the granitic or continental layer; (*e*) possibility of the elimination of surface waves by locating the detonating and recording points on opposite sides of a steep-walled canyon; (*f*) presence or absence of a fault and its accurate location; (*g*) dip of the fault surface.

Field experimentation was carried on at five localities in southern and central California: two localities in Owens Valley, and one each in Yosemite Valley, the Ventura Basin, and the Los Angeles Basin.

The authors gratefully acknowledge the aid received from Dr. William S. W. Kew, geologist of the Standard Oil Company at Los Angeles, in furnishing information and advice regarding the geology of the Los Angeles and Ventura basins.

Instruments and Apparatus

The assembly of instruments and accessory apparatus used in these experiments is designed for the seismographic registration, at several station-points separated by suitable distances, of earth vibration due to the passage of elastic waves generated by explosion, usually, at a more or less distant point, called the "shot-point," whence they are propagated through the materials of the upper part of the earth's crust, along direct and reflected or refracted paths. The prime object of this registration is to determine with sufficient accuracy the intervals between the time of the explosion, or other originating cause, and the times of arrival of the direct and reflected or refracted waves at the places where the detecting instruments are placed.

In the present experiments the earth disturbances in all cases were produced artificially by shallow explosions of very small to small magnitude.

Theoretically, especially for very complete or precise studies, instruments to register all three components of vibratory motion at each station-point would be required and the constants of the seismometer systems would need to be known. However, much can be learned if only the time intervals between the explosion and the arrivals of identifiable waves can be determined accurately, together with the wave-speeds, and this can be found from the records of a single component without exact knowledge of all the constants of the seismometer. It was this information which was sought in these experiments and vertical-component detectors only were used.

Six detecting elements of the seismographic assembly were used, at six different station-points, usually along a straight line through the shot-point. These detectors are small, completely encased electro-magnetic mechanisms ruggedly constructed to withstand the necessary handling and transportation in the field. In use they could be, and were, buried in shallow holes to minimize the effects of wind, grass movement, and other disturbances at the ground surface.

Essentially, these detecting elements consist of an inertia-mass suspended (vertically) so as to have a short free period of vibration, mounted in a closed cylindrical case, filled with oil to provide sufficient damping. Exposure to earth vibration produces relative motion of the mounting and the inertia-mass, which actuates an electro-magnetic device so as to produce electromotive forces which vary with the vibration. The oscillating electric current thus generated is conducted to a galvanometer element and thereafter registered photographically on a moving band of photographic paper or film.

Thus the varying response of each detecting element is led to the central station (mounted on a motor truck, which carries the registering apparatus and photographic equipment and supplies, and also serves to carry wire reels, tools, supplies, and explosives for field use) along an electric circuit which usually consists of a single wire with ground return but sometimes is a complete two-way metallic circuit. Because of their necessarily rugged construction the direct sensitiveness of the detecting elements is not great. For the same reason the intrinsic sensitivity of the galvanometer elements is not high. Consequently the electric currents which reach the central station from the detectors are very feeble, and it is necessary to amplify them to operate the galvanometer system satisfactorily. The amplifiers used are of special design so that the galvanometer elements actuated by their output currents record vibrations within a certain range of frequency, say 10 to 1,000 cycles, with much greater magnification than those of higher or lower frequency. In other words, the assembly has been designed and adjusted to be specially effective for the specific purpose of recording the artificially generated waves it is used to detect.

The string type of galvanometer was used in the present case, but a much less sensitive form than the usual laboratory string galvanometer, as it was used with currents from the output of an amplifier. To facilitate registration, a harp type of assembly was employed, with six strings placed parallel and close together, each connected through its amplifier circuit with one of the six detecting elements referred to above. With a suitable optical assembly for illumination, magnification, and focusing, the point-shadow of each of these strings, determined by a very narrow slit at right angles to their length, was registered on the moving photosensitive surface so as to form six parallel continuous lines along the band of recording paper or film. In the absence of disturbance these lines are straight. When the detectors are disturbed by the passing waves, or otherwise, the lines become zigzag or sinuous, with excursions of the shadow-points to one side and the other of the mean or rest position.

For determining time intervals, time lines are marked on the record at right angles to its length and direction of motion. These lines are spaced at intervals of one-hundredth of a second, with distinguishing lines of greater strength every tenth of a second. There are several ways in which these time lines may be registered. In all, essentially, the light which illuminates the slit, at right angles to the galvanometer strings (which thus determine the point-shadows of the strings) is eclipsed for a brief instant every one-hundredth of a second, leaving a very narrow unexposed strip which, on development, becomes a line at right angles to the lines formed by the shadows of the strings. In registration on paper, or the direct record on film, the time lines and the lines of the strings are light lines on a dark ground. Of course, if contact photographic copies are made from the film records, black lines on a white ground result in the usual way.

Provision is made for the immediate development, fixation, and washing of the photographic records in suitable light-trapped tanks, and the records are available for study within a few minutes after registration.

In field practice, generally, the explosive is buried in a hole, usually of slight depth, dependent on the amount of the explosive and other circumstances. In some cases the charge is placed directly on the surface. Wet or damp ground renders the explosion far more effective than dry, and the shock produced is especially strong when it is possible to fill the hole with water.

The charge is connected by wire with the firing mechanism, placed at a short, safe distance from the shot-point. The firing mechanism serves to detonate the explosive after all is in readiness and the recording mechanism has been tested and set in motion. To insure proper co-operation between the operators at the firing-point and at the central station, a telephone circuit is provided which enables them to communicate with each other. This circuit, sometimes a single wire with ground return, serves also to connect the firing mechanism with the recording galvanometers, so that the instant of operation of the mechanism, which is also (nearly enough) the instant of explosion, is recorded directly, usually on one string but sometimes on more. In fact the circuit constants are so arranged that the instant indicated on the record is the instant as closely as it is reasonably possible to determine it.

There are also arrangements for changing resistances so as to diminish the responses, and the resulting amplitudes, for the direct waves near the beginning of the record. This sometimes is necessary to prevent illegibility due to too great recorded amplitudes, or to prevent the breaking or entanglement of the galvanometer strings.

THEORETICAL CONSIDERATIONS

General theory.—If at some point of a body the equilibrium is disturbed by a sudden stress, a change in volume and a shear are propagated through the body. If the point of disturbance is close to the surface, surface waves also are propagated. The theory may be found in any book on elasticity or modern seismology.[3] If the disturbance is an explosion, the shear caused in the body will be small as compared with the change in volume, and in this case most of the energy will be propagated in the form of compressional waves. Indeed, transversal waves (shear waves) have not been observed during the investigations described in the following pages. In fact, to observe shear waves, in this case commonly having periods of one-tenth of a second or more, instruments (especially for the horizontal components) are required which have considerable magnification for waves with periods of this order. Surface waves are generated either by the immediate action of shear and change in volume close to the surface—this is the most important mode of origin of surface waves—or when elastic waves reach a surface. Surface waves, moreover, consist of waves with relatively long periods (more than one-twentieth of a second) in the case of the

[3] For example, B. Gutenberg, "Theorie der Erdbebenwellen," "Handbuch der Geophysik," **4**, Verlag Gebr. Borntraeger, Berlin, 1929

explosion of dynamite just below the surface. In our investigations no normal surface waves were observed, as the instruments (vertical components only) did not have large magnification for waves with periods of one-tenth of a second and more. Long waves, registered in a few cases, may have been surface waves caused by longitudinal or transversal waves encountering the surface of the earth or discontinuities at depth. These will be dealt with later. Neglecting them for the present, compressional waves (longitudinal waves) only which were propagated either through the soil or through the air were observed. The velocity of such waves in a medium with the density *d*, the bulk modulus *k*, and the rigidity μ is given by

$$V^2 = \frac{k + \frac{4}{3}\mu}{d} \tag{1}$$

The rigidity of air is practically zero, and the bulk modulus under normal conditions is proportional to the density and the absolute temperature, T. In this case, with sufficient approximation,

$$V_{\text{air}} = 20.1\sqrt{T} \tag{2}$$

During these investigations the temperature in general was around 30° C. (86° F. or 303° abs.). In this case we find from Equation 2 that the velocity of sound is about 350 meters per second. If the air is in motion we must add to the value calculated from Equation 2 the velocity of the air in meters per second along the direction from the shot-point to the instrument. If the travel-time in the air is known ("travel-time" means the time which the wave requires to travel from the origin to the point of observation), we can calculate the distance of the instrument from the shot-point. This method was often used; in some cases, however, the distances were measured. If the sound method was used, a small charge (a small fraction of a pound of dynamite) was exploded on the surface, and the air waves were registered. If it was expected that the waves through the air would arrive at a time when no waves through the earth were due, two charges were exploded simultaneously, one at some depth to cause elastic waves in the surface layers of the earth's crust, and one on the surface to cause air waves. In general a buried charge did not set up air waves large enough to be registered by the instruments, and usually the waves excited in the earth by charges put on the surface were not large enough for study.

Unfortunately the use of the sound method for finding the distance of the shot-point from the instruments becomes inaccurate with increasing distance[4] because of the fact that the paths of the sound waves and their apparent velocity along the surface will change whenever there are differences in temperature or wind conditions or both at different elevations, as usually occurs. The error found by Gabriel[4] generally was less than two per cent of the distance up to 350 meters.

The fundamental problem is to cause an explosion at some depth, to find the travel-times of the elastic waves which, in the following pages, are always supposed to be longitudinal waves, and to calculate from these data the velocity of the waves as a function of geographical longitude, latitude, and depth. Another problem will be to determine the material traversed by the waves along their course. Theoretically, the first can always be accomplished by using different quantities of explosives, different shot-points, and different sorts of waves (direct, reflected, refracted). Therefore it is always possible theoretically to locate discontinuities between materials of different wave-velocities. However, it is not possible to determine the nature of the material at depth at a given place, for a large number of different materials may have the same velocity. In general, knowledge of the geology or information from bore holes will determine what material is present in a special case, or geological considerations will limit the possibilities.

The theory of the procedure as a whole may be divided into two parts: the theory of physical action at the instant of the explosion, and the theory of the propagation of the waves. (The recording of the waves is a technical problem which has been treated in the preceding section.) Up to date the second part only, the propagation of the waves, has been treated in detail.

The energy propagated by the waves depends upon the kind and quantity of explosive, the depth to which it is buried, and the material in which the explosion takes place. Not only the amount of energy produced by a given quantity of the explosive, but also the speed of explosion affects the energy radiated into the earth. Slowly exploding materials, such as black powder, produce seismic waves of considerably

[4] Compare, for example, V. G. Gabriel, "Some Experience in Seismic Prospecting," *Gerlands Beitrage zur Geophysik,* Erganzungshefte fur angewandte Geophysik, **2**, 122, 1931.

smaller amplitudes than rapidly exploding materials. The explosives used for geophysical prospecting usually have a speed of explosion of about seven kilometers per second, or somewhat less. In most cases blasting gelatine with 10–80 per cent nitroglycerin is used. The following investigations were carried through with 60 per cent Hercules gelatine. The maximum charge was sixty pounds, but sometimes only an electric blasting cap was exploded, without dynamite. As stated above, the charge was put on the surface of the ground when records of the sound waves were desired. In this case a considerable amount of energy goes into the air. On the other hand, if the charge is buried so deep that it does not blow out when exploded, the air waves are very weak. When it is desired to have the maximum amount of energy enter the earth—even in this case the greater part of this generated energy is wasted in the development of heat—the charge must be placed so deep that the explosion does not blow out. The larger the charge is, the deeper must be the hole. As only hand-drilled holes, usually not more than a few feet deep, were available for this work, it was necessary to divide the larger charges among two or three holes, and this also tended to limit the quantity of dynamite used in any single explosion.

The material in which the charge is detonated has a large effect on the energy transmitted into the earth. If the dynamite is exploded in very dry sand, consisting of relatively large grains, a great part of the energy will be lost in compacting the mass. The material will act like an assemblage of springs, and relatively long waves with small energy will be propagated through the earth. The less porous the material is in which the dynamite is exploded, the more energy will be radiated as elastic waves. Thus very good seismograms were obtained when the explosive was detonated in water. Dynamite, exploded in a hole in granite covered by water to a depth of a very few feet, generated waves of relatively large amplitudes in the granite, so that charges of a fraction of one pound could be utilized.

Around the cavity in which the explosion occurs, the stresses are generally greater than the crushing strength of the material, and fracture results. But within a small distance a zone is reached where the movements are purely elastic. There the displacement begins with the impulse of the initial wave, followed in general by a few waves caused by repeated oscillations of the particles near the origin. As seen above porous sands behave essentially as springs and only longer waves are set up in them. Therefore, there are differences in the movements even at the beginning of the propagation of the waves. But still larger differences arise owing to the internal friction in the material. Without internal friction the waves and impulses would be propagated with unaltered forms and periods, but as soon as internal friction is involved, all wave forms flatten, impulses disappear, and the periods of individual waves increase. The theory of the propagation of disturbances in a medium with internal friction shows[5] that in the primary equations the coefficient of rigidity μ must be replaced by the operator

$$\mu\left(1 + C\frac{\partial}{\partial t}\right) \tag{3}$$

in which C is a constant of the material, and t the time. The bulk modulus should not be changed, as in the case of pure compression the particles do not move past each other and therefore no internal friction is caused If the Lamé constant λ is used, it must be replaced in the primary equations by the operator

$$\lambda - \tfrac{2}{3}\eta\frac{\partial}{\partial t}, \text{ in which } \eta = \mu C; \quad C = \frac{\eta}{\mu} \tag{4}$$

For metals η (the "coefficient of internal friction") is of the order of 10^9 dyn-sec./cm.^{2} and C of the order of 0.001 sec. K. Sezawa[6] has derived general equations for compressional waves in a plane. Gutenberg[7] found that in general in such a case the change of period of an individual wave from the value T_0 seconds at the origin to the value T at a distance D is given to a certain approximation by

$$T^2 = T^2_0 + aD \tag{5}$$

in which a is a constant, depending upon the material. a is inversely proportional to V^3 (V = wave-velocity) and in addition depends upon the values of μ and η, mentioned above.

It is very difficult to compare observations with this theory, as the records at different distances are usually obtained from different shots

[5] B. Gutenberg and H. Schlechtweg, "Viskositat und innere Reibung fester Korper," *Physikalische Zeitschrift,* **31**, 745, 1930.

[6] K. Sezawa, "On the Decay of Waves in Visco-Elastic Solid Bodies," *Bulletin of the Earthquake Research Institute,* Tokyo, **3**, 43, 1927.

[7] B Gutenberg, "Handbuch der Geophysik," **4**, 22, 1929.

and usually also from different shot-points. Hence the waves in different seismograms will have different periods. By using average values from the same region, however, this effect will be eliminated to a certain extent. More serious are the facts that the instruments do not reproduce with the true period the first waves of a wave-train, or the waves of a wave-train in which the periods change with time, and that the magnification of waves with different periods is different. It is especially true that instruments with galvanometric recording have a maximum magnification for waves of periods within a small range, and the magnification of shorter and especially of very much longer waves decreases very considerably. Nevertheless an attempt has been made to determine from our records whether a certain average relation exists between the distance from the point of explosion to the instrument and the period of the waves recorded. The six instruments in general extended along a line not exceeding 300 meters, but usually less. Their records showed no noticeable difference in the period registered in the case of any particular explosion. Especially, when the nearest instrument was installed at a distance between forty and fifty meters from the shot-point, the longitudinal waves were registered usually with a period of about 0.01 of a second and the most distant instrument (150 to 200 meters) recorded no periods which differed noticeably from this value. On the other hand, in the case of the longest distances, which unfortunately for our purpose were all in the same region (Los Angeles Basin), up to eight kilometers, the periods of the first longitudinal waves ranged from 0.02 to 0.04 of a second. As an average, the following values may be given, but it must always be taken into consideration that they are somewhat uncertain for the reasons given above:

Distance in meters..............	50	200	1000	8000
Period in thousandths of a second.	10	12	20	30

These values follow Equation 5 within the limits of error, if a has the value 0.0001 and D is measured in kilometers. The calculated values in this case are:

Distance in meters...............	0	50	200	1000	8000
Period in thousandths of a second..	10	10	12	14	30

The agreement extends to the fact that in a circle of up to about 150 meters from the point of explosion the period of the waves seems to be constant. Some idea as to how this period originates has been given above, but this problem needs further investigation.

The effects of the instruments on the records have been mentioned above. There is still the question how the ground on which the instruments stand affects the results. As a first approximation we can consider the ground as a more or less damped seismograph. Thus, if the damping is small (water-filled mud or sand), free oscillation will begin after the first waves have passed on, but otherwise no large effect will arise in this way.

As Equation 1 shows, the velocity of the waves in a homogeneous medium depends upon the elastic constants of the medium and its density. Now the question arises whether the wave velocity can be different in the same material for waves of different periods. Theory[8] shows that absorption may cause dispersion; long waves are somewhat delayed. No noticeable dispersion has been observed in the case of longitudinal earthquake waves up to date. It is unreasonable that longitudinal waves produced by explosions should show noticeable dispersion so long as the wave-length is not large as compared with the thickness of the layers, in which case the ordinary theory is void. No detailed theoretical investigations seem to have been made for this case.

The surface waves on the other hand very often show dispersion. They are propagated close to the surface with the velocity corresponding to the uppermost layers. But what is to be considered as "uppermost layers" depends upon the wave-length. In the case of earthquakes the wave-length varies between a very few and more than a thousand kilometers. In the first case the "uppermost layer" which must be considered is a few kilometers thick and the velocity of surface shear waves in it is 3.2 kilometers per second or even less. In the second case the "uppermost layer" has a thickness of hundreds of kilometers, and the average velocity of the surface shear waves is of the order of 4½ kilometers per second. Similar effects must be produced in the case of artificial surface shear waves, but, as we have seen, these types of waves could not be found in the records of our instruments. Therefore, in the following considerations it may be supposed that the velocity of the waves used for our purposes and similar ones does not depend upon the wave-length, so long as the opposite has not been found from experiments.

Travel-time curves and discontinuities.—The theory of propagation of elastic waves may be found in any modern book on seismology. The fundamental law used in applied seismology connects the angles of in-

[8] Compare for example B. Gutenberg. "Handbuch der Geophysik," **4**, 25, 1929.

cidence i_1 and i_2 at any two points (for example E and A in Figure 1) of the same ray and the velocities V_1 and V_2 at these two points by the equation

$$\frac{\sin i_1}{\sin i_2}=\frac{V_1}{V_2}\text{, or, in general, } \sin i_1 : \sin i_2 : \sin i_3 \ldots = V_1 : V_2 : V_3 : \ldots \quad (6)$$

All formulas concerning travel-times needed in applied seismology can be derived from this formula and purely geometric considerations.

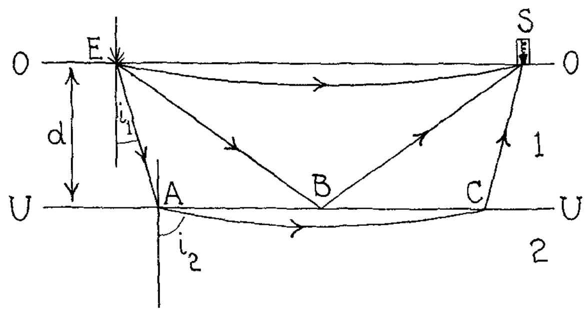

FIG. 1

We will now suppose that we have a region (Figure 1) with a layer 1, in which the velocity V_1 increases slightly with depth, and a layer 2 with the velocity V_2 noticeably larger than V_1, also increasing slightly with depth, as usual; such increase of velocity with depth is to be expected because of the increase of pressure. No supposition is made as to the reason why the velocity increases suddenly at the surface UU, which we suppose to be horizontal like the surface OO. There may be a change in material, or layer 1 may be formed by dry sand, and layer 2 by the same sand below the water table, or there may be a change in the size of grains of the same material. If a charge is exploded at the point E, longitudinal waves will travel along the direct paths E to S, where S is the observing station (instrument). If the distance is not too great, the downward curvature of the rays caused by the slight increase in velocity with depth will have no noticeable effect on the travel-time t. If D is the distance E to S, we find therefore

$$t = \frac{D}{V_1} \quad (7)$$

If, on the other hand, we observe the travel-times of waves caused at E by explosions at different distances, plot the travel-times against the distances, and find that this "travel-time curve" is a straight line through the origin, we may conclude that we have observed the longitudinal wave through the uppermost layer, and that the velocity there is given by

$$V_1 = \frac{D}{t} \tag{8}$$

Another wave should be observed at S, caused by a wave reflected at B, traveling along EBS. From purely trigonometric considerations we find the travel-time of the reflected wave:

$$t = \frac{1}{V_1} \sqrt{4\,d^2 + D^2} \tag{9}$$

We see that the travel-time curve begins horizontally with $t_0 = \frac{2d}{V_1}$ and approaches the travel-time curve of the direct waves at large distances, as Equation 9 changes into Equation 7 as d becomes very small as compared with D. If we observe two reflected waves at two distances D_1 and D_2 (travel times t_1 and t_2), we find

$$V_1 = \sqrt{\frac{D^2{}_2 - D^2{}_1}{t^2{}_2 - t^2{}_1}} \qquad d = \tfrac{1}{2}\sqrt{V^2{}_1\, t^2{}_1 - D^2{}_1} = \tfrac{1}{2}\sqrt{V^2{}_1\, t^2{}_2 - D^2{}_2} \tag{10}$$

Figure 1 shows that in the supposed case there is a third wave, traveling from E to S, over the path $EACS$. This is the refracted wave through the deeper layer. The angles i_2 and i_1 are connected by Equation 6. This refracted wave, contrary to the wave considered previously, cannot be observed at all distances. From Equation 6 the following equation may be derived:

$$\sin i_2 = \frac{V_2}{V_1} \sin i_1 \leqq 1, \tag{11}$$

so that the refraction cannot take place, if sin i_1 is larger than $\frac{V_1}{V_2}$. If this "critical angle" i^* is exceeded, no refracted wave is generated. For the critical angle itself the refracted wave is the same as the reflected wave, which means that the travel-time curve of the refracted wave begins at a certain point on the travel-time curve of the reflected wave, which, according to Equation 9, is always later than the direct wave.

If the angle i_1 is slightly smaller than the critical angle, i_2 will be close to 90°, and the path of the refracted wave between A and C will be nearly the same as the straight line ABC. Again the wave will travel on a slightly curved path caused by the increase of velocity with depth on account of pressure, as Formula 6 shows that the rays become curved as soon as the velocity changes, either gradually or suddenly. To find the order of the effect of the increase of velocity with depth on waves leaving a discontinuity nearly horizontally, it may be supposed that the rays are circles. In this case formulas derived by E. Wiechert[9] can be used in a simplified form, as the depth d in our case is always very small compared with the radius r_0 of the earth. The velocity V in such a case can be written in the form $V = \frac{1}{2} f^2(A^2 - r^2)$, in which $r = r_0 - d$. Thus it may be supposed, neglecting d^2 with respect to $r_0 d$, $V = \frac{1}{2} f^2 (D + 2r_0 d)$, in which $D = A^2 + r^2_0$. If, on the other hand, we suppose that the velocity is given by $V = a + bd$, we find

$$D = \frac{2ar_0}{b} \qquad\qquad f^2 = \frac{b}{r_0}$$

Wiechert found that a constant C, given by $C = \dfrac{A^2 + r^2_0}{A^2 - r^2_0}$, or, in our case, as D is small compared with r^2_0, $C = \dfrac{2r^2_0}{D} = r_0 \dfrac{b}{a}$, is of importance in the theory. The distance $\Delta°$, at which the ray, leaving the surface at an angle of incidence i, returns to the surface, and this angle i itself are connected by the equation $\cot i \cot \Delta = C$, or, in our case: $(90° - i) = \frac{1}{2}C\Delta$, in which Δ and i are measured in degrees. If we measure Δ in kilometers, we have $i = 90° - 0.0045C\Delta$ kilometers. Finally, we calculate the maximum depth h of the ray from the surface. In our case by use of Equation 6 it is found that $h = \dfrac{a}{b}(1 - \sin i)$.

To get numerical results, we must make a supposition as to the increase of velocity with depth. Investigations on the increase of the bulk modulus with pressure, carried out at the Geophysical Laboratory of the Carnegie Institution of Washington and elsewhere, as well as observations, have shown that the rate of the increase of velocity with depth in a practically homogeneous medium may be of the same order

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[9] E. Wiechert, "Ueber Erdbebenwellen I," *Gottinger Nachrichten,* 1907; B. Gutenberg, "Handbuch der Geophysik," **4,** 36.

as that to be expected from the effect of pressure. This increase depends very much upon the material. It is especially high in porous rocks and generally is larger in rocks with low than in rocks with high bulk modulus under normal conditions. If we suppose that we have, just beneath a discontinuity near the surface of the earth, a velocity of 2 00 kilometers per second, and in the same layer at a depth of 100 meters beneath the discontinuity, a velocity of 2.05 kilometers per second, these values correspond to observed velocities, as well as to observed changes in the bulk modulus due to pressure. In this case we have $V = 2 + \frac{1}{2}d$, in which all quantities are given in kilometers and seconds: therefore $a = 2$, $b = \frac{1}{2}$, $C = 1{,}600$, $i = 90° - 7\Delta$ km, $h = 4(1 - \sin i)$. We find the following angles of incidence (i), and deepest points (h), beneath the discontinuity for rays which have run a distance Δ in the deeper layer:

Δ (meters)	100	500	1,000	2,000
i	89°.7	86½°	83°	76°
h (meters)	0.3	7½	30	120

These results, which give merely an idea of the form of the rays which leave the surface practically horizontally, show that at small distances the rays always run very close to the surface. This result may explain the fact that observations on explosion-waves as well as records of near earthquakes very often do not show the refracted waves clearly at the shortest distances where they should be observed, but very much better at longer distances. If the wave, which we call "refracted wave," were caused by a longitudinal surface wave traveling along the discontinuity, as has been supposed by some, just the opposite should be expected: we should find large waves at the shorter distances and a noticeable decrease in intensity with increasing distance, as the theory shows that the energy of the longitudinal surface wave decreases very rapidly with distance

Even if the velocity increases very much more slowly with depth than we have supposed, at longer distances the waves penetrate very deep into the second layer; h is proportional to $1 - \sin i$, in our case $1 - \sin i = \frac{1}{2}\cos^2 i = \frac{1}{2}\sin^2(90° - i)$, and, as $(90° - i_0)$ is proportional to $C\Delta$, we see that h is proportional to $C^2\Delta^2$, which means to $\left(\frac{a}{b}\right)^2$ and to the square of the distance. If, for example, the thickness of the layer were 100 meters, the direct wave could not reach a distance of 2,000 meters, but would be reflected or refracted earlier at the deeper surface.

At greater depths in the earth the velocity increases very much more slowly with depth. Investigations concerning the bulk have shown that, at pressures exceeding 1,000 atmospheres, the bulk modulus increases very much more slowly, and at pressures over 2,000 atmospheres, corresponding to a depth of around seven kilometers in the earth, no noticeable increase of the bulk modulus with increasing pressure has been found. We do not know very much about the increase of velocity with depth in this region. In the mantle of the earth at depths between 100 and 1,200 kilometers the value of the constant C is around seven, but even this may be caused partly by change in material.

For calculations concerning blasts or local earthquakes, we always can assume within the limits of error that AC (Fig. 1) is a straight line. The idea mentioned above that a surface wave is propagated along AC and that EA and SC are perpendicular to the surface UU has been suggested, as in some cases the refracted wave arriving at S seemed to have no horizontal component. But investigations on these waves with sensitive instruments have shown that there is a small horizontal component in the refracted wave arriving at S, so that the idea of the vertical return of the refracted wave has been proved to be wrong, and to have been based on records of instruments not sensitive enough for such investigations.

As we have seen, the refracted wave begins at a certain distance, and the corresponding angle of incidence is the critical angle i^* given by

$$\sin i^* = \frac{V_1}{V_2} \tag{12}$$

With increasing distance, i_1 will decrease and the center of the path AC will remove more and more from the straight line AC, but for all calculations in applied seismology it may be supposed within the limits of error that i_1 is always equal to i^* and that AC is a straight line. Thus we may suppose for purposes of calculation that the whole path of the refracted wave consists of the straight lines EA, AC, and CS, where

$$EA = CS = \frac{d}{\cos i^*} \text{ and } AC = D - 2d \tan i^*$$

(D = distance ES). Therefore the travel-time of the refracted wave is given by

$$t = \frac{2d}{V_1 \cos i^*} + \frac{D - 2d \tan i^*}{V_2}$$

If we introduce, according to Equation 12,

$$\cos i^* = \sqrt{1 - \left(\frac{V_1}{V_2}\right)^2}$$

and

$$\tan i^* = \frac{V_1}{V_2 \cos i^*}$$

we find finally

$$t = 2d \sqrt{\frac{1}{V^2_1} - \frac{1}{V^2_2}} + \frac{D}{V_2} = a + \frac{D}{V_2} \tag{13}$$

in which a does not depend upon the distance D,

or

$$d = \frac{V_2 t - D}{2\sqrt{\frac{V^2_2}{V^2_1} - 1}} \tag{14}$$

If we observe the travel-times t_1 and t_2 of the refracted wave at the distances D_1 and D_2, we find from Equation 13

$$t_2 - t_1 = \frac{D_2 - D_1}{V_2} \qquad V_2 = \frac{D_2 - D_1}{t_2 - t_1} \tag{15}$$

If we have three layers with horizontal boundaries and the velocities V_1, V_2, and V_3, we may suppose that in the third layer the angle i_3 is 90°, and by use of Equation 6 we find $\sin i_1 \,.\, \sin i_2 : 1 = V_1 : V_2 : V_3$. The further procedure is now very similar to that in the preceding case, and similar relations may be applied if we have to deal with four or more layers with horizontal surfaces.

When, as usual, the velocity V_2 is higher than the velocity V_1, the travel-time curves of the direct wave ES and the wave $EACS$ refracted through the second layer will intersect at a distance D^*, which is given from Equations 7 and 13 by

$$\frac{D^*}{V_1} = a + \frac{D^*}{V_2} \text{ or } D^* = \frac{a}{\frac{1}{V_1} - \frac{1}{V_2}} \text{ in which } a = 2d\sqrt{\frac{1}{V^2_1} - \frac{1}{V^2_2}} \tag{16}$$

from which we may derive

$$d = \frac{D^* \left(\frac{1}{V_1} - \frac{1}{V_2} \right)}{2 \sqrt{\frac{1}{V^2_1} - \frac{1}{V^2_2}}} \tag{17}$$

If we have n layers with horizontal boundaries, at very short distances we shall observe waves through the first layer only. From their travel-time curve we can derive the velocity V_1 in this layer by means of Equation 8. At somewhat larger distances we find the waves refracted through the second layer. The velocity V_2 there can be found from the travel-time curve of these waves from Equation 15. In a similar way the velocities V_3, V_4, and so on are to be found from the travel-time curves of the waves with their deepest point in the third and fourth layers, and so on.

The thickness of the first layer is given by Equation 14 or Equation 17. To find the thicknesses of the deeper layers, the following equations, derived like the preceding formulas, may be used:

$$\sin i_1 : \sin i_2 : \sin i_3 : \ldots : 1 = V_1 : V_2 : V_3 : \ldots : V_n \tag{18}$$

$$D_n = 2d_1 \tan i_1 + 2d_2 \tan i_2 + 2d_3 \tan i_3 + \ldots + 2d_{n-2} \tan i_{n-2} \tag{19}$$

$$t_n = \frac{2d_1}{V_1 \cos i_1} + \frac{2d_2}{V_2 \cos i_2} + \frac{2d_3}{V_3 \cos i_3} + \ldots + \frac{2d_{n-2}}{V_{n-2} \cos i_{n-2}} \tag{20}$$

If t' is the travel-time at a distance D' of the wave with its deepest point in the layer n, we calculate

$$X_n = D' - D_n \qquad T_n = t' - t_n \tag{21}$$

$$d_{n-1} = \frac{(V_n T_n - X_n) \cos i_{n-1}}{\left(\frac{V_n}{V_{n-1}} - \frac{V_{n-1}}{V_n} \right)} \tag{22}$$

We begin with $n = 3$ and find the thickness d_2 of the second layer, then we repeat the whole procedure with $n = 4$ to find the thickness of the third layer and so on.

Up to this point it has always been supposed that the surfaces are horizontal. If they are still planes, but not horizontal, the formulas are more complicated. Methods to find the thicknesses of such layers, and the velocities, have been given for example by E. A. Ansel[10] and by O. von Schmidt,[11] but we shall not use such formulas in the following investigations.

Energy and Amplitudes.—The preceding formulas connect the thicknesses of the layers and the wave-velocities with the travel-times of the different waves. But the relative amplitudes are also important in many cases. We suppose that the energy radiates homogeneously in all directions from the source. It will decrease on account of absorption and may undergo other physical changes. We may suppose as a first approximation that this loss of energy depends on the distance only. Moreover it is small as compared with the effect of the increasing wave front and with other factors which we shall now consider. As soon as a longitudinal wave arrives at a boundary between two regions, in which at least one of the velocities (longitudinal or transversal) or the density of material change, the energy of the wave spreads into four new waves: two refracted and two reflected waves, in each layer one longitudinal and one transversal. Sometimes one or the other may not appear, if the critical angle considered above has been passed. Formulas to compute the ratio of energy transmitted into these four waves have been given by Knott,[12] and formulas to compute the amplitudes by Zoeppritz.[13] H. P. Berlage[14] has given some methods for calculating approximative values.

To reproduce the equations of Zoeppritz, we denote the ratios of the amplitudes by the following letters:

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[10] E. A. Ansel, "Das Impulsfeld der praktischen Seismik in graphischer Behandlung," *Gerlands Beitrage zur Geophysik, Ergänzungshefte für angewante Geophysik,* **1**, 117, 1930.

[11] O. von Schmidt, "Theorie der 3-Schichten-Seismik," *Zeitschrift fur Geophysik,* **7**, 37, 1931.

[12] C. G. Knott, "Reflexion and Refraction of Elastic Waves, with Seismological Applications," *Philosophical Magazine,* **48**, 64, No. 290, July 1899.

[13] K Zoeppritz, "Ueber Erdbebenwellen VII b," *Nachrichten der Koniglichen Gesellschaft der Wissenschaften* zu Gottingen, mathematische-physikalische Klasse, 1919, S 57. (Published twelve years after the death of Zoeppritz.)

[14] H. P. Berlage, "Naherungsformeln," *Gerlands Beitrage zur Geophysik,* **26**, 131, 1930.

$$\left.\begin{array}{l} l = \text{ratio longitudinal reflected to incident amplitude} \\ L = \text{ratio longitudinal refracted to incident amplitude} \\ t = \text{ratio transversal reflected to incident amplitude} \\ T = \text{ratio transversal refracted to incident amplitude} \end{array}\right\} \begin{array}{l}\text{All positive in the direction of} \\ \text{propagation } (l \text{ and } t \text{ into me-} \\ \text{dium 1, } L \text{ and } T \text{ into 2)}\end{array}$$

We denote the corresponding angles of incidence, reflection, and refraction by a, b, c, d, in which a is also the angle of incidence of the arriving longitudinal wave; if furthermore the velocities are given by V_1 (longitudinal in layer 1, where the wave arrives), v_1 (transversal in layer 1), V_2 (longitudinal, layer 2) and v_2 (transversal, layer 2), if the densities in these two layers are d_1 and d_2, and if finally we use the following abbreviations:

$$\left.\begin{array}{c} F = \dfrac{V_1}{v_1} \quad G = \dfrac{d_2}{d_1}\left(\dfrac{v_2}{v_1}\right)^2 \dfrac{V_1}{V_2} \quad H = \dfrac{v_1}{V_1} \quad I = \dfrac{d_2}{d_1}\dfrac{V_2}{V_1} \\ K = \dfrac{d_2}{d_1}\dfrac{v_2}{V_1} \qquad J = \dfrac{d_2}{d_1}\left(\dfrac{v_2}{v_1}\right)^2 \dfrac{V_1}{v_2} \end{array}\right\} \quad (23)$$

the equations found by Zoeppritz can be written in the following form·

$$\left.\begin{array}{r} \sin a + l \sin a - t \cos c - L \sin b + T \cos d = 0 \\ \cos a - l \cos a - t \sin c - L \cos b - T \sin d = 0 \\ -\sin 2a + l \sin 2a - Ft \cos 2c + GL \sin 2b - JT \cos d = 0 \\ -\cos 2c - l \cos 2c - Ht \sin 2c + IL \cos 2d + KT \sin d = 0 \end{array}\right\} \quad (24)$$

in which according to Equation 6

$$\sin a : \sin b : \sin c : \sin d = V_1 : V_2 : v_1 : v_2.$$

If the ratios of the velocities and the ratio of the densities are given and therefore the values of the constants F $(= 1/H)$, G, I, J $(= GI/K)$ and K, the amplitudes l, L, t, and T can be calculated for every angle a.

Knott has given formulas for the energies. He used the following abbreviations:

C = cotangent of the angle of incidence of the arriving longitudinal (and the reflected longitudinal) wave

C' = cotangent of the angle of incidence of the refracted longitudinal wave

γ = cotangent of the angle of incidence of the reflected transversal wave

γ' = cotangent of the angle of incidence of the refracted transversal wave

n = modulus of rigidity in layer I (arriving wave)

n' = modulus of rigidity in layer II (refracted waves)

A = energy factor of the arriving longitudinal wave

A_1 = energy factor of the reflected longitudinal wave

A' = energy factor of the refracted longitudinal wave

B_1 = energy factor of the reflected transversal wave

B' = energy factor of the refracted transversal wave

$$X = A + A_1 \qquad Y = A - A_1 \tag{25}$$

Then he finds the following equations:

$$\left.\begin{aligned} B_1 + cY &= B' + c'A' \\ \gamma B_1 + X &= -\gamma' B' + A' \\ -2\gamma B_1 + (\gamma^2 - 1)\,X &= 2\frac{n'}{n}\gamma' B' + \frac{n'}{n}[(\gamma')^2 - 1]A' \\ (\gamma^2 - 1)\,B_1 - 2cY &= \frac{n'}{n}[(\gamma')^2 - 1]B' - 2\frac{n'}{n}c'A' \end{aligned}\right\} \tag{26}$$

which, together with Equation 25, give the values of A_1, A', B_1, and B' as a function of A_1 of the elastic constants n and n' and the wave-velocities for a given angle of incidence.

The energy equation then is given by

$$cd_1A^2 = cd_1A_1^2 + \gamma d_1B_1^2 + c'd_2\,(A')^2 + \gamma' d_2\,(B')^2 \tag{27}$$

in which d_1 is the density in the first and d_2 the density in the second medium On the left-hand side of the equation we have the energy of the incident longitudinal wave, on the right-hand side the energies of the reflected and refracted waves. The ratio of the energy E_f of the refracted longitudinal wave to the energy of the incident wave, and the energy E_r of the reflected longitudinal wave, for example, are therefore given by the following equations (if E_i = energy of the incident longitudinal wave):

$$\frac{E_f}{E_i} = \frac{c'd_2(A')^2}{cd_1A^2} \qquad \frac{E_r}{E_i} = \frac{A^2_1}{A^2} \tag{28}$$

Calculations have been carried through in a few cases by Knott[15] and in others by Slichter and Gabriel.[16] One of the cases considered by Knott is based on values, which may occur occasionally, to a certain approximation, in applied seismology. He supposed that the materials of the two layers have the same density (in general they do not differ very much; exceptions occur in the case of ice, coal, and other unusually light or heavy materials), and that the velocities have the ratios

$$V_1 : V_2 : v_1 : v_2 = 1\ 82 : 2.05 : 1.00 : 1.31$$

The results will be the same, if all values are changed in the same ratio Knott found in this case the following results:

The reflected longitudinal wave receives 4 per cent of the energy when the incident wave arrives at a right angle at the surface. With increasing angle of incidence a the energy of the reflected longitudinal wave decreases, reaches a minimum of 0.2 per cent of the incident energy corresponding to $a = 15°$ *ca.*, and remains rather weak up to $a = 60°$. For $a = 62\frac{1}{2}°$ the critical angle of the refracted longitudinal wave is reached, and in the case of waves arriving at a greater angle of incidence, more than two-thirds of the energy passes into the reflected longitudinal wave. At distances, where the refracted longitudinal wave exists, this carries nearly all the energy. The refracted transversal wave is, in general, small, and its maximum energy of 20 per cent occurs near the critical angle of the refracted longitudinal wave. But in general this wave is of no interest for applied seismology, as at the points where the longitudinal wave encounters surfaces on its way down, the refracted

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[15] See footnote 12.

[16] *Gerlands Beitrage zur Geophysik, Erganzungshefte fur angewandte Geophysik*, abstract in *Bulletin of the American Physical Society*. Vol 6, No 7, December 11, 1931

transversal wave with its lesser velocity has a smaller angle of incidence and pitches down from the surface so steeply that it does not return to the region where registration of the waves takes place. On the other hand the longitudinal waves coming back from the depth may set up, at surfaces, some weak transversal waves, but they follow so soon after the longitudinal waves that usually they cannot be separated from longitudinal waves traversing similar paths. Finally, the refracted transversal waves, in the case calculated by Knott, do not receive one per cent of the energy up to an angle of incidence of 60°. At about this angle they carry a maximum of 13 per cent of the energy, but all transversal waves are practically zero for longitudinal waves arriving nearly parallel to the surface of the earth. In this case, and in others mentioned above, the refracted longitudinal wave carries by far the greater part of the energy; the reflected longitudinal wave receives some energy in the case when the incident longitudinal waves arrive nearly vertically at the surface; with increasing angle of incidence it decreases nearly to zero, but it receives a much larger part of the energy as soon as the critical angle for the refracted wave has been passed. The transversal waves can be neglected in the case of artificial explosions.

If the distance of the instruments from the point of explosion is small as compared with the thickness of the layers, the longitudinal waves arrive nearly vertically. In this case all angles in the System 24 are nearly zero. If we suppose vertical incidence, it follows from the first and the third equation of the System 24 that both transversal waves (t and T) are zero; from the second and last equation we find

$$l = \frac{I-1}{I+1} \qquad L = \frac{2}{I+1} \qquad \text{in which } I = \frac{d_2 V_2}{d_1 V_1}\,. \tag{29}$$

l is the amplitude of the reflected and L of the refracted longitudinal wave. The reflected wave will be the larger, the more I differs from 1. If, for example, the densities in two layers are equal, and the velocities of the longitudinal waves have the ratio 2 : 1, then I is 2, and the amplitude of the reflected wave is one-third of the amplitude of the arriving wave.

Equations 26 and 28 of Knott in the case of a longitudinal wave with vertical incidence become:

$$\frac{A_1}{A} = \frac{I-1}{I+1} \qquad \frac{A'}{A} = \frac{V_2}{V_1}\frac{2}{(I+1)} \tag{30}$$

$$\frac{E_r}{E_i} = \frac{(I-1)^2}{(I+1)^2} \qquad \frac{E_f}{E_i} = \frac{4I}{(I+1)^2} \tag{31}$$

If, again, we calculate the values corresponding to $I = 2$, we see that one-ninth of the arriving energy is reflected In one and the same medium the energy is proportional to the square of the amplitudes and inversely proportional to the square of the periods.

These changes in energy at surfaces of discontinuity are not the only major causes which affect the final amplitudes. A certain amount of energy leaving the origin between two certain rays in some cases will arrive at the surface within a small area; in others it will be spread out very much. In the first case every unit of the surface receives very much more energy than in the second one. If, for example, we rotate the three rays, leaving the point E in Figure 1, by the same amount, the end of the direct ray will move very far, the end of the reflected ray will move very much less, the end of the refracted ray very much more. We see that (neglecting now the purely physical effect of the discontinuity) every unit of the surface will receive the most energy from the reflected wave, less from the direct wave, and still less from the refracted wave. Other causes for differences in energy at different points are the facts that as the circumference of the area on which the energy arrives increases with distance, the energy density decreases, and that the flux of energy reaches the surface at different angles of incidence. The ratios of the amplitudes[17] at a certain distance Δ are given to a first approximation by $A\sqrt{F \tan i \frac{di}{d\Delta}}$ (leaving out a factor $\sqrt{1/\Delta}$, common to all) in which for the horizontal component

$$A_{\text{hor}} = a\frac{\cos\alpha}{\cos 2\alpha} \qquad a = \frac{1.-M}{1+M} \qquad M = \frac{\tan i}{\tan\alpha}\cot^2 2\alpha \qquad \frac{\sin\alpha}{\sin i} = \frac{V_{\text{trans}}}{V_{\text{long}}}$$

and for the vertical component

$$A_{\text{vert}} = A_{\text{hor}} \cot 2\alpha.$$

F = product of the successive ratios of reflected or refracted energy to the incident energy at the successive points of reflection or refraction

[17] For details of all these effects see · B Gutenberg, "Handbuch der Geophysik," 4, 57.

$i =$ angle of incidence of the wave under consideration at the station

$\Delta =$ distance

A depends only upon the ratio of the wave-velocities (longitudinal : transversal) and the angle of incidence at the earth's surface. In general the following values may be used:

i	0°	5°	10°	20°	30°	40°	50°	60°	70°	80°	85°	90°
A_{hor}	0.00	0.19	0.39	0.75	1.08	1.36	1.57	1.66	1.61	1.26	0.83	0.00
A_{vert}	2.00	1.99	1.97	1.86	1.71	1.50	1.27	1.04	0.81	0.54	0.34	0.00

In the direct wave F is equal to 1, i is close to 90°, tan i is large, $\frac{di}{d\Delta}$ is very small (as we have seen by rotating the ray) and A depends largely upon i. In the refracted wave i is usually in the median numerical range, tan i is of the order 1, A is of the same order, $\frac{di}{d\Delta}$ is small, and F is close to 1, as the refracted waves carry nearly all the energy. In the case of the reflected wave F is very small, only a very small percentage, as we have seen. If the instrument is close to the shot-point, i is only a few degrees, tan i is small and A_{hor} is very small, while A_{vert} is practically 2. Supposing again a horizontal reflecting plane, the order of $\frac{di}{d\Delta}$ can be found from the equation $\tan i = \frac{\Delta}{2d}$, in which $d =$ depth of the reflecting discontinuity, $\frac{di}{d\Delta} = \frac{\cos^2 i}{2d}$. To get an idea of the values which occur, we suppose that the curvature of the direct and the refracted waves is given by the data found on page 200. We find in this case the following orders for the quantities at distances of 50 to 100 meters:

Direct wave

$i = 89°.7 \qquad \frac{di}{d\Delta} = \frac{0.3}{100}$ degrees per 100 meters $\qquad A_{hor} = 0.1$ (*ca.*)

$A_{vert} = 0.04$ (*ca.*) $\qquad \tan i = 200$ (*ca.*)

Refracted wave:

$\sin i = \frac{V_0}{V_d} \qquad V_0 =$ Velocity near surface of the earth, $V_d =$ Velocity at the deepest point. If $\frac{V_0}{V_d} = \frac{1}{2}$, so $i = 30°$, $\tan i = 0.58$, $A_{hor} = 1$,

$A_{\text{vert}} = 1.7$. If the amplitudes are reduced to nine-tenths by the refraction, we have $F = 0.9$.

Reflected wave:

If $\Delta : 2d = 0.05$, we have $\tan i = 0.05$, $i = 3°$, $A_{\text{hor}} = 0.1$, $A_{\text{vert}} = 2$. If, besides, only one per cent of the incident amplitudes is carried by the reflected wave, we have $F = 0.01$. The calculation gives the following results: the horizontal amplitudes of the direct, the refracted, and the reflected waves are to each other approximately as 1: 0.5: 0.1; the vertical components are in the ratios as 1 . 3 : ⅓. If the reflecting plane has a certain dip, this must be considered, as it affects the angles of incidence of the reflected and refracted waves and therefore also A, F, and $\frac{di}{d\Delta}$.

These results, of course, are only intended to show how the calculations are made, and how the different factors affect the result.

At the surface the energy is divided into components by the instruments. If, for example, we have two horizontal seismographs, one recording motions in the direction through the shot-point, and one perpendicular to it, the latter instrument will register no longitudinal waves at all. The former will record nearly the entire amplitude of the direct wave, a certain fraction of the reflected wave, especially from distant sources, and a small fraction of the refracted wave. On the other hand, as we may see from our calculations above, a vertical seismograph will record only a small fraction of the direct wave, a large part of the reflected wave, especially near the origin, and a certain fraction of the refracted wave. For this reason vertical component seismographs are used chiefly in applied geophysics.

Refraction and reflection methods.—The results developed in the preceding pages show that there are two different ways of using the records of explosions for finding the wave-velocities and the discontinuities in the earth's crust. We may utilize either the refracted waves (refraction method) or the reflected waves (reflection method). The first investigator who seems to have recognized that it is possible to use such methods at all appears to have been Mallet.[18] A general description of the fundamental ideas of the refraction methods was given by Belar [19] A. Sieberg[20]

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[18] R Mallet, *The Transactions of the Royal Irish Academy,* **21,** 1848.

[19] A. Belar, "Eine neue praktische Verwendung der Erdbebenmesser," *Die Erdbebenwarte,* **1,** 1901, 1902.

[20] A Sieberg, "Handbuch der Erdbebenkunde," p. 333 (Braunschweig, 1904).

was the first to write a special chapter on "applied seismology." In later years, E. Wiechert published his fundamental ideas on the propagation of earthquake waves and showed how to find the velocity if the travel-time curve is known. In 1908 he suggested the use of artificial shocks for finding the structure of the outer parts of the earth's crust exactly as had been done with earthquake waves, which means that he was the first to suggest clearly the details of the refraction method. He also encouraged Mintrop to investigate in Göttingen the elastic waves caused by falling weights, and many years later Mintrop[21] was the first to succeed in using the method for practical purposes. Today the refraction method is usually employed. From the records of explosions the travel-times of the direct and the refracted waves are calculated, the travel-time curves are traced, and then the Formulas 16 to 22, the methods of Ansel or von Schmidt, or similar procedures are used to find the velocity of the longitudinal waves as a function of position and depth.

The first to suggest and use reflection of waves for prospecting seems to have been R. A. Fessenden.[22] He used sound waves and detectors. The method as it is used now, with seismographs and oscillographs, was first suggested by B. Gutenberg.[23] So far as we know, the first to use it successfully in the field appears to have been J. C. Karcher. The waves caused by an explosion arrive at a few seismographs connected with the different elements of an oscillograph. (This arrangement of course is advantageous in the case of the refraction method, too, where usually only one seismograph is used, and on the other hand the reflection method might use one instrument only, but this would cause much loss of time.) On the records the reflected waves will arrive at very much shorter time intervals than the direct or refracted waves, especially near the origin, as may be seen from Equation 9. If the average velocity in the upper layer is known (either from information from wells in the neighborhood or from a few refraction shots), and reflections are seen on the seismograms, the depth of the reflecting surface can be found at once. If it is possible to use reflections at different distances from the point of explosions, the Formula 10 will give all necessary results In general the slope of the reflecting surface will not affect the result noticeably. In

[21] Deutsches Reichs Patent 371963, December 7, 1919; United States of America Patent 1,599,538, September 14, 1926.

[22] R. A. Fessenden, "Method and Apparatus for Locating Ore-Bodies" (U.S.A. Patent 1,240,328, 1917).

[23] B. Gutenberg, "Lehrbuch der Geophysik," pp 609, 610 (Gebruder Borntraeger, Berlin) The part containing this method was published in 1926.

case of a steeply sloping surface, circles drawn around the different points of observation with the observed distances to the reflecting surface as radii will indicate the location and dip of the surface.

In general the reflection method has been used very little in prospecting; Mothes and others have used it to find the thickness of ice (glaciers in the Alps and in Greenland). It has been feared that the energy of the reflected waves is inadequate. Indeed, only a very small percentage of the energy is reflected at a discontinuity in the case of practically vertical incidence of the arriving ray, but the flux of energy is very much more favorable to great amplitudes in this case than in any other one as we have seen. On the other hand, at long distances from the source, when the critical angle of the refracted wave is exceeded, by far the greater part of the arriving energy is reflected, but this energy is now spread over a very large area of the surface at a long distance from the origin, and consequently the amplitudes of the reflected waves are not very large.

Both the reflection and the refraction methods have advantages and disadvantages, and our investigations have shown that sometimes it is advisable to use them together. If it is desired to trace a certain discontinuity through a considerable area and the depth of this discontinuity is known at one point, the reflection method offers many advantages over the refraction method, as by reflection every successful observation gives at once the distance of the discontinuity from the shot-point, even if the reflecting surface is not a plane. If the refraction method is used a series of explosions and observations must be made at least twice along a line with the recording instruments first at one and then at the other end of the line, which, besides, must be of considerable length, if the discontinuity is at a considerable depth. If, on the other hand, investigations are made in a region in which nothing is known about the structure, it is desirable first to make refraction observations to find the velocities at different depths. The following results of our investigations will show better than any description the use of the two methods, their advantages and disadvantages.

Velocities in different rocks.—The seismic methods give us the position of discontinuities and the wave velocities. The final problem is, now, to determine the materials constituting the different layers. In many cases geologists will be able to recognize them by using the known values of the velocity of longitudinal waves in different rocks. In Table I velocities of longitudinal waves are given from observations. In special cases the values in different samples of the same kind of material may differ still more than is shown in the table, as composition, porosity, and water content affect the wave velocity notably.

TABLE I

VELOCITY OF LONGITUDINAL WAVES

	km/sec.		km/sec		km/sec.
Gravel, sand, dry	½–1	Limestone	1½–2	Salt . .	4½–5¼
Wet loam .	¾–1	Sandstone .	2–2½	Granite .	4¾–5¾
Wet sand . . .	¾–1¼	Schist	3.1±	Basalt . . .	5–6
Sandy clay	1–1¼	Chalk . .	3¼–4		
Water . .	1.4–1.5	Ice .	3½		

ALABAMA HILLS FAULT AND THICKNESS OF SEDIMENTS

Geology.—For the seismic investigation of faults two localities were chosen, one near Lone Pine and the other near Shepard Creek, both in the southern part of Owens Valley. This depression is one of the most profound structural troughs on the North American continent. Its length is over 100 miles, its width is from three to ten miles, and its depth is nearly two miles. It is a great graben bounded on both sides by faults or fault zones, above which rise the imposing scarps of the Sierra Nevada on the west and the Inyo Mountains on the east. The floor of the valley has an altitude of about 4,000 feet; the scarp on the west rises to a maximum height of 14,496 feet in Mount Whitney. In the southern part of Owens Valley the block the depression of which formed the graben is split by a fault or fault zone into two parts. The western part has been tilted westward and the uplifted eastern margin of it presents an escarpment looking eastward over the other half of the main block. This uplifted eastern margin forms the Alabama Hills.

One of the two localities in Owens Valley at which experimentation was carried on lies at the east base of these hills, west of the highway, and about three miles south of Lone Pine. The area utilized was the then dry bottom of the intermittent Diaz Lake and the meadow south of it.

The rocks constituting the Alabama Hills immediately to the west are, as shown on Knopf's[24] excellent reconnaissance map of the region, metamorphosed Triassic shales and pyroclastics. These rocks still show bedding distinctly and are not as highly crystalline as the intrusive rocks of the region. Diaz Lake basin and meadow are underlain by soft lake

24 "United States Geological Survey Professional Papers," No. 110, 1918

beds, the deposits of the expanded Quaternary Owens Lake. Covering these is a mantle with presumed thickness of a few tens of feet of later alluvial fan, soil, and Diaz Lake deposit differing but little from the underlying uncompacted Quaternary Lake beds.

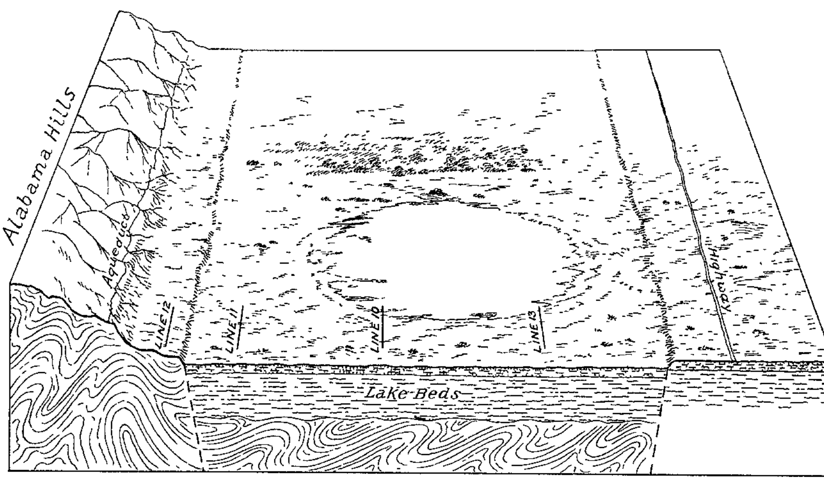

FIG. 2—Plat of Diaz Lake and meadow, three miles south of Lone Pine, California, showing fault structure, relation of sediments to old crystalline rocks, and location of lines of shot-points. Generalized view, looking north

Diaz Lake and meadow lie on a depressed north-south strip of ground somewhat less than one-half mile in width. Both escarpments originated by recent movement on the two faults which bound the depressed strip or miniature graben. These faults are branches of the fracture zone on which a vertical displacement occurred in 1872, causing the severe Owens Valley earthquake. The east-facing escarpment west of Diaz Lake is not a part of the main east-facing scarp of the Alabama Hills, which lies some hundreds of feet farther west; the two scarps are separated by a flat terrace cut by radial swinging of one of the intermittent streams issuing from the hills. This bedrock terrace permitted determination of velocity in the old rocks immediately west of the western fault. The west-facing scarp on the east side of Diaz Lake reveals only soft lacustral sediments. While the position of the two faults can be quite accurately located on the ground by the escarpments, neither is exposed in section, and their dips at this locality are not known. The thickness of the soft lacustral sediments in the Diaz Lake basin, resting on the Triassic recrystallized rocks, was likewise not ascertainable by ordinary geologic means.

Seismic investigations.—Two days were spent in investigating discontinuities near the fault at the eastern edge of the Alabama Hills. Four profiles were shot parallel to the eastern base of the hills. The lines 10, 11, and 13 were on the alluvium east of the hills at distances of 210, 46, and 600 meters, respectively, from the base, and the Profile 12 was on a sloping terrace at a distance of three meters west from the base

TABLE II

TRAVEL-TIMES OF LONGITUDINAL WAVES, PROFILE 12

(In thousandths of a second)

Distance in meters ...	56	77	97	107	133	160	143	164	184	194	220	245
Time in thousandths of a second........	64	71	79	85	95	104	94	99	104	110	119	123
Difference in thousandths of a second.	−2	−2	0	3	5	4	1	−1	−2	1	2	1

The figures derived from the records of the two explosions on line 12 are given in Table II. In Figure 4 the travel-times are plotted against the distance. The travel-time of the longitudinal waves along the line between distances of 60 and 250 meters is given by

$$t = 0.049 + \frac{\Delta}{3.1} \text{ sec.},$$

in which Δ equals the distance from the shot-point to the station, measured in kilometers. The differences between the calculated and the observed values are given in the last line of Table II. The minus sign means that the observed values are earlier than the calculated values. Both agree within the limits of error, though instead of the number 3.1 in the formula another with a second decimal may give still better agreement. The formula has the type of Equation 13 indicating that we have a thin layer with low velocity (alluvium), which velocity is not determinable from our figures, but which must be less than or equal to $56:64 = 0.9$ kilometers per second (very probably close to that number), and that beneath the alluvium is a layer with a velocity of 3.1 kilometers per second (crystalline gneiss and schist of the Alabama Hills).

The profile closest to the east base of the hills, but on the alluvium,

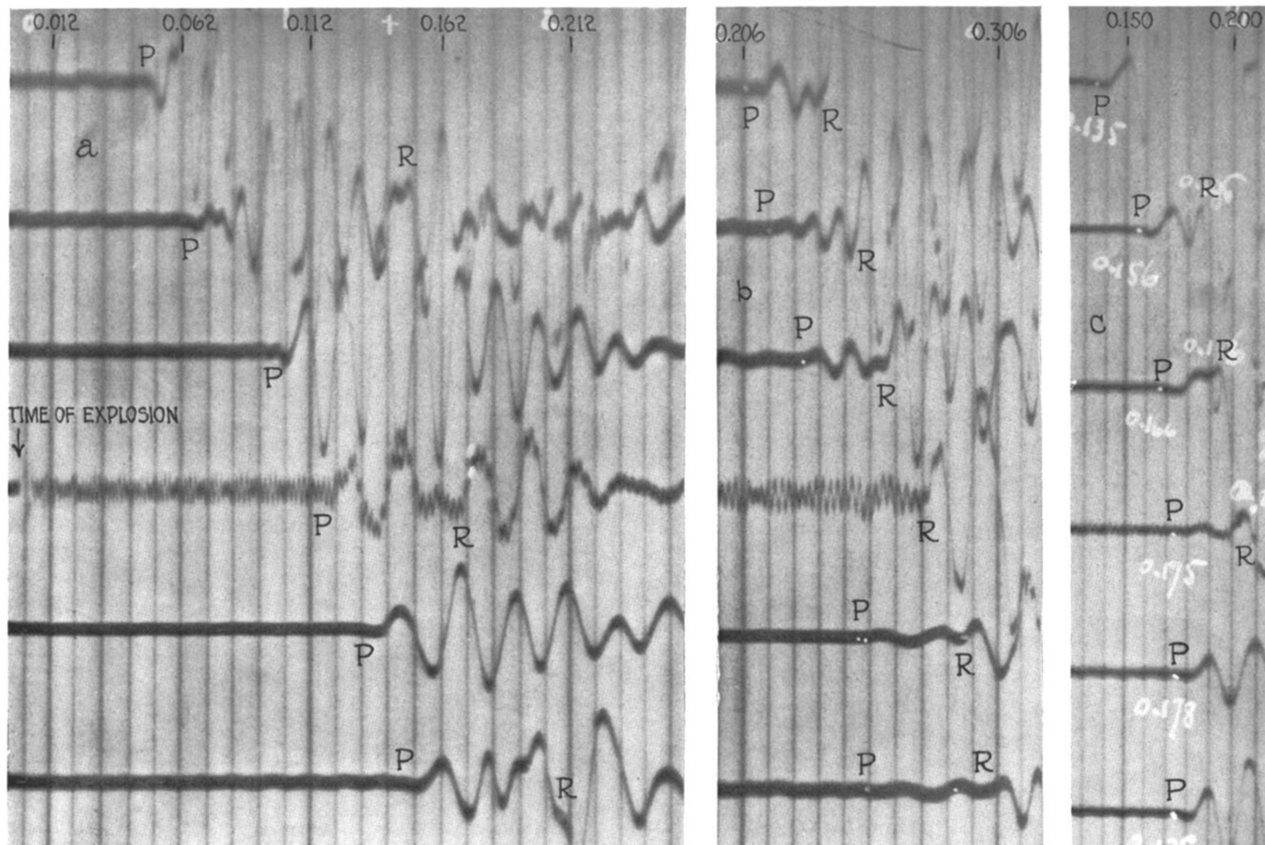

FIG. 3.—Seismograms: (*a*) from shot-point 11 (east of Alabama Hills), distances of the instruments from the shot-point 70 to 221 meters, charge, ½ pound of dynamite, (*b*) from shot-point 11, distances 333 to 481 meters, charge, 6 pounds of dynamite, (*c*) from shot-point 2 (east of Sierra fault), distances 136, 153, 168, 184, 202, and 223 meters, charge, ½ pound of dynamite. The vertical lines indicate the time at intervals of 1/100 of a second, the figures at the top show the time since the explosion in seconds

is line 11. As has been stated, it is parallel to the base of the hills at a distance of 46 meters. Four shot-points were used, while the six instruments remained at the same points. Copies of records obtained at the first and the third of these shot-points are reproduced in Figures 3*a* and 3*b* respectively. In most cases there is one very noticeable phase, marked *R* in the seismograms, after the beginning *P*. The times of both kinds of waves have been measured, and plotted against the distances in Figure 4 (p. 218).

It may be recognized that there are three curves marked by *a, b,* and *c* in the figure. *a* and *b* correspond to the phase *P* in the seismograms, *c* to the phase *R*. The travel-times of these waves are given in Table III.

TABLE III

TRAVEL-TIMES OF WAVES,* PROFILE 11 (4 SHOT-POINTS)

Curve *a*			Curve *b*			Curve *c*		
Distance	Time	Difference	Distance	Time	Difference	Distance	Time	Difference
70	45	—2	388	239	2	100	149	—1
100	63	—3	452	261	4	158	170	5
221	148	9	481	266	0	221	194	6
196	122	—2	479	266	1	254	200	—1
225	146	5	512	272	—3	348	240	—4
254	163	4	597	306	4			
288	185	5	630	312	—1	333	247	10
315	204	8				363	258	7
348	211	—5				388	271	7
						422	286	6
333	216	9				452	301	6
363	225	0				481	316	6
						479	310	1
						512	322	—4
						534	336	—2
						568	357	0
						597	376	5
						630	384	—5

* Distance in meters, time and difference in thousandths of a second.

The travel-time t corresponding to a straight line connecting the points of Curve a is given by $t = 0.005^{s} + \frac{\Delta}{1.65}$ in which $\Delta =$ distance in kilometers. The straight line corresponding to b is given by

$t = 0.117^s + \dfrac{\Delta}{3.1}$. The differences "observed minus calculated values" are given in Table III. Both equations are of the type of Equation 13, so that we may conclude that the two lines *a* and *b* correspond to the travel-times through two layers in which the velocities are 1.65

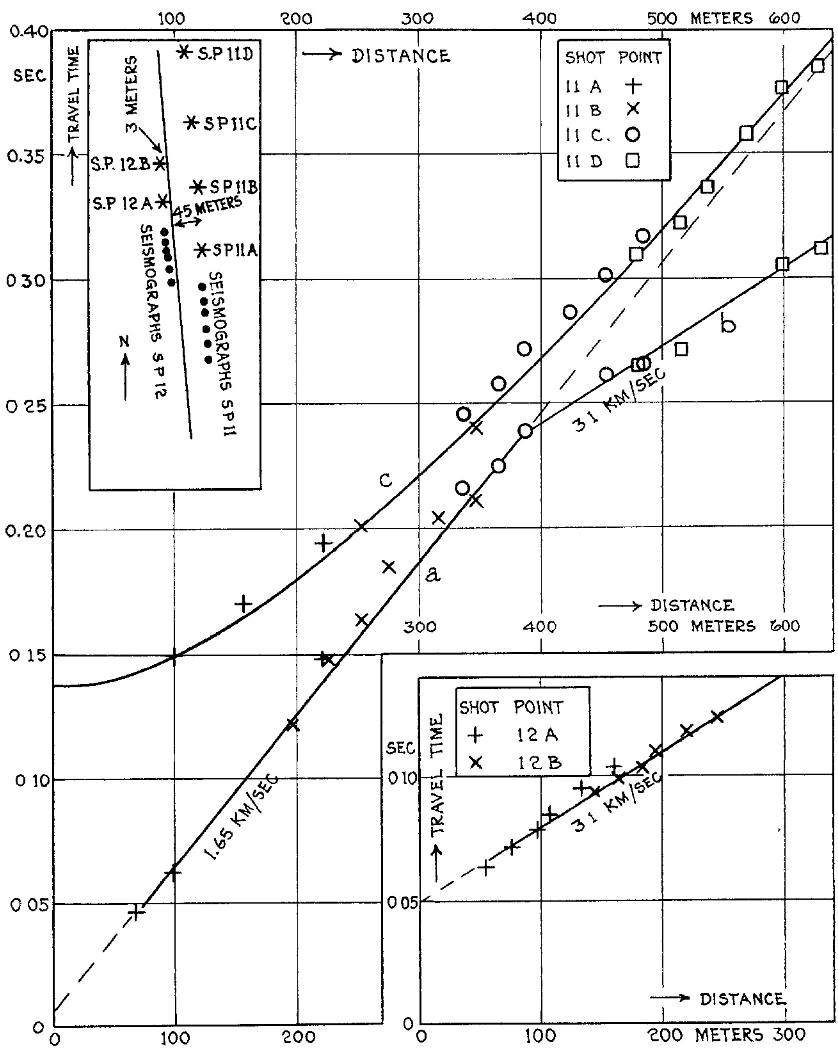

FIG 4.—Chart showing travel-times near Alabama Hills.

and 3.1 kilometers per second, supposing that the boundary between both is nearly horizontal. No reason has been found in this case to doubt this supposition. The first layer corresponds to the alluvium in the valley near the hills. It is covered by a few feet of surface material in which the velocity of longitudinal waves is less than 1 65 kilometers per second, which leads to the constant term 0.005 of a second in the first equation The denominator 1.65 in this equation shows that the velocity of longitudinal waves in the alluvial layer is 1.65 kilometers per second. In the same way we see that in the deeper layer the velocity of longitudinal waves is 3.1 kilometers per second. As this value is just the same as the value found from Profile 12 for the material of the Alabama Hills, it is very probable that the waves whose arrival is given by the travel-time Curve *b*, run through the crystalline schists of the Alabama Hills. To decide whether this material is at a certain depth beneath the alluvium of the valley, or whether we have records of waves which were running to the west into the Alabama Hills, crossing the fault, and then again into the alluvium and to the instruments, we calculate the distance to the surface at which the refraction takes place by using Formula 17, neglecting the effect of the few feet of sand near the surface. The two lines intersect at $D^* = 395$ meters, and so we find

$$d = \frac{D^* \left(\frac{1}{v_1} - \frac{1}{v_2} \right)}{2 \sqrt{\left(\frac{1}{v_1} \right)^2 - \left(\frac{1}{v_2} \right)^2}} = 110 \text{ meters}$$

As the horizontal distance of the profile from the base of the Alabama Hills was only 46 meters, it follows that the refracting surface is at depth, and not a vertical or nearly vertical plane through the base of the hills.

If we now consider Curve *c*, we find that it has the properties of a travel-time curve for reflected waves; for very small distances it begins at a certain time and afterwards approaches the travel-time Curve *a*. If this supposition is correct we may apply Equation 10 to calculate the velocity of the waves in the layer and its thickness. If we use the first point of Curve *c* and the average of the last two points, we find in this way a thickness of 119 meters and a velocity of 1.73 kilometers per second. Both values correspond to those found before from the Curves *a* and *b*, within the limits of error, especially if we consider that the waves, corresponding to Curve *a*, have run in the upper part of the alluvium only, but that the reflected waves *R* have passed through the whole alluvium layer, in which the velocity of waves very probably increases somewhat with depth. The equation of the travel-time curve of the reflected waves corresponding to the values mentioned is $t = \frac{\sqrt{\Delta^2 + 0.057}}{1.73}$. The differences, "observed minus calculated values," are given in the last column of Table III. They are somewhat larger than in the preceding cases (maximum 1/100 sec.), partly due to the fact that their beginning is sometimes less clear than the beginning of the first phase, and partly to the fact that the assumption of a horizontal plane reflecting surface is probably not fulfilled.

The following Profile 10 runs parallel to the base of the hills at a distance of 210 meters. Two shot-points were used, the position of the instruments being the same for both. Again reflected waves were registered, but no refracted waves, corresponding to the branch *b* of the travel-time curve belonging to shot-point 11. The travel-times are given in Table IV. The differences, "observed minus calculated values," refer

TABLE IV

TRAVEL-TIMES OF LONGITUDINAL WAVES,* PROFILES 10 AND 13

Profile 10						Profile 13		
Direct Waves			Reflected Waves			Direct Waves		
Distance	Time	Difference	Distance	Time	Difference	Distance	Time	Difference
44	28	—5	364	251	0	46	36	—4
60	36	—5	380	259	1	76	57	1
77	47	—5	397	268	—3	92	68	0
94	64	2	412	275	1	107	80	3
110	74	2	428	284	2	122	85	—1
364	221	5						
380	230	1						
397	241	—1						
428	260	—3						

* Distance in meters, time and difference in thousandths of a second

to the same equations that have been used in the preceding table Thus it seems to be very probable that at a distance of 200 meters from the base we still have the crystalline material of the hills at a depth of somewhat over 100 meters, though waves through this layer have not been observed, since the profile was not long enough for this purpose. Finally one charge was fired on line 13, running parallel to the base of the hills at a distance of 600 meters. Waves through the alluvium only were registered. Their travel-times are given by $t = 0.072^s + \frac{\Delta}{1.65}$.

The results found from the investigations near the Alabama Hills have been treated first, since they show very clearly the direct, the refracted, and the reflected waves, and because the results derived by use of the reflection and refraction methods agree very well. The travel-time curves of Profile 11 are outstanding examples in illustration of the theory.

The experiments show besides that it is possible in this way to get information as to the structure near faults in a very few days, if the alluvium layer near the fault has a thickness not exceeding a few hundred meters. Examples of thick layers will be given later.

Sierran Fault and Sediments near Shepard Creek, Owens Valley

Geology.—A second locality for experimentation in Owens Valley was chosen along the east side of the main Sierran fault, about one mile south of the mouth of Shepard Creek and about eight miles southwest of Independence. It is northwest of the Diaz Lake locality. Huge coalescing alluvial fans extend for about six miles with gradually increasing slope and coarseness of materials, from the middle of Owens Valley up to the base of the 10,000-foot granitic east-facing scarp. At the Shepard Creek locality the fault lies not at the base of the main escarpment, which has receded about a half-mile to the west, but at the base of a fresh minor escarpment thirty or forty feet in height, also facing eastward. Granitic rock forms the low escarpment and the terrace above it. Knopf characterizes the rock as quartz monzonite, including rocks ranging from granite to quartz diorite. At the experimentation locality some parts of the mass are gneissic.

While the base of the escarpment locates the fault quite closely, the fracture cannot be seen and its dip is not evident. The thickness of the alluvial fan, the head of which at this locality stands about 2,000 feet above the floor of Owens Valley, is not apparent, but it would appear to be measurable in hundreds at least and quite possibly in thousands of feet While the surface portion of the fan at the Shepard Creek locality consists of ordinary soil, the underlying materials are coarse gravels including many huge bowlders with diameters measurable in feet.

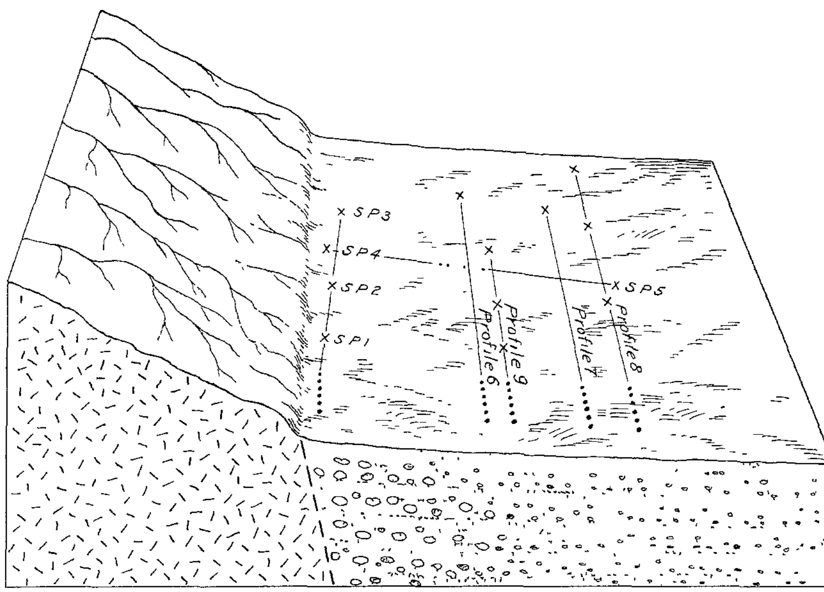

FIG 5.—Generalized view, looking north, along east base of Sierra Nevada eight miles southwest of Independence, California, indicating relations of Sierra scarp, fault, alluvial fan, and the profiles or lines of shot-points along which reflection and refraction measurements were made

Seismic investigations.—Three days were spent in investigating the structure close to the fault zone at the base of the Sierra on the western side of Owens Valley north of Shepard Creek. In the following, the word "fault" is always used to indicate the line where the fault appears to be according to signs at the surface.

The profiles 1, 2, and 3 were parallel to the fault, the position of the instruments being the same in all three cases. The distance between the instruments and the fault was twenty-five meters (see sketch at the upper left of Figure 6). The shot-points were at different distances from the instruments (Table V, p. 224). The travel-times (Figure 6 and Table V) show that the velocity in the alluvial layer is close to 1.0 kilometers per second, and that this covers a layer with very high velocity. As granite is visible on the western side of the fault it is very probable that this high velocity is attributable to a granite layer. In this case a velocity of 5 to 5½ kilometers per second should be expected. No certain value can be derived from the very short curves. From the

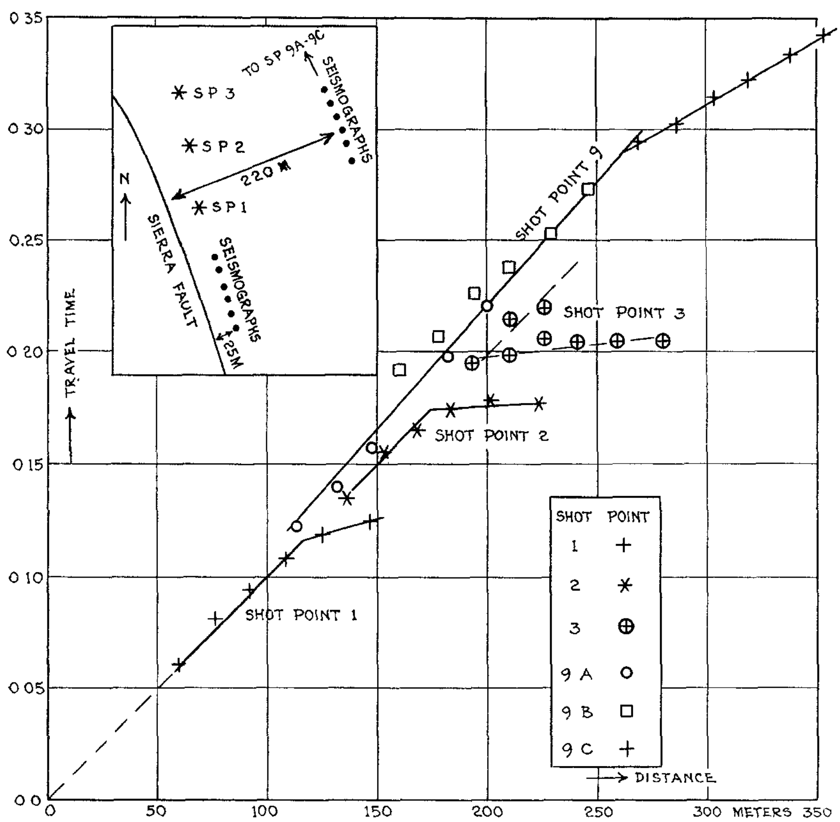

FIG 6.—Chart showing travel-times near Shepard Creek.

travel-time curves 1 to 3 in Figure 6 it is clear that shot-point 3 was more distant from the granite than shot-point 2 and still more distant than 1. No attempt has been made to find the direction of the refracting plane As soon as the seismograms of the first shot had been developed, it was clear that there is indeed a discontinuity. Figure 3*c* shows a seismogram from shot-point 2. The distances between the different instruments were of the same order, but time intervals between the beginnings on the seismograms differ very much, indicating a well-marked change in velocity in the neighborhood. So there is no doubt that faults between layers with very different elastic constants can be discovered by use of one explosion and can be traced accurately with only a small expenditure of time

TABLE V

Travel-Times of Longitudinal Waves* near the Sierra Fault in Owens Valley

Shot-Point 1		Shot-Point 2		Shot-Point 3		Shot-Point 4		Shot-Point 5	
Distance	Time	Distance	Time	Distance	Time	Distance	Time	Distance	Time
60	60	136	135	194	195	51	52	61	64
76	82	153	156	210	199 214	102	94	159	202
92	94	168	166	226	207 220	150	146	211	274
108	108	184	175	242	205	198	197	258	310
125	119	202	178	259	206	250	248	312	327
147	125	223	176	280	206	348	330	360	362

* Distance in meters; time in thousandths of a second

The question now arises whether the refraction took place near the fault or at a certain vertical depth. If the first is true, the distance of the shot-point 3 from the fault was the greatest; in the second case, the thickness of the sediments increases very rapidly toward the north. To decide this question, the instruments were put along a line perpendicular to the direction of the fault, and two shots were fired, number 4, at a point between the fault and the instruments and twenty meters east of the fault, and number 5, at a point distant 430 meters from the fault, with the instruments between shot-point and fault. No sign of waves with high velocity was found in either case (Table V). Some irregularities in the arrival time may have been caused by the very heterogeneous composition of the ground material.

The following profiles, 6 and 7, were parallel to the fault at distances of 200 meters and 350 meters respectively. Again they showed no traces of refracted or reflected waves. The same holds in the case of Profile 8 (parallel to the fault at a distance of 370 meters) and Profile 9 (parallel at a distance of 220 meters; see sketch in Figure 6). In both three shot-points were used and the profiles were extended over a considerable distance. Both travel-time curves (for shot-points 9 in Figure 6) give an average velocity of about one kilometer per second, up to a distance of 260 meters (Profile 9) and 350 meters (Profile 8) respectively, and about 1½ kilometers per second for distances up to

550 meters (Profile 8). In both cases the observed travel-times scatter around the average values. They seem to indicate that the thickness of the sediments is greater than 250 meters and that the refraction in the profiles 1, 2, and 3 very probably is caused by the boundary between the alluvium and the granite west of the fault. No reflected waves were observed in the seismograms of the profiles 4 to 9, indicating again a great thickness for the sediments east of the Sierra fault. In all these cases the waves generated were very long, the period being up to one-twentieth of a second.

TABLE VI

TRAVEL-TIMES OF LONGITUDINAL WAVES,* PROFILES 8 AND 9

Profile 8						Profile 9					
Distance	Time	Distance	Time	Distance	Time	Distance	Time	Distance	Time	Distance	Time
196	168	321	297	408	363	114	122	160	194	269	295
226	200	350	333	442	403	131	140	177	209	286	303
257	233	381	361	467	410	148	158	194	224	303	315
287	300	412	404	493	420	182	198	209	239	318	323
317	324	440	404	524	456	199	226	228	254	337	334
346	352	470	412	552	470			245	274	354	343

* Distance in meters, time in thousandths of a second.

INVESTIGATIONS IN YOSEMITE REGION

Physiography and geology.—Yosemite Valley is the glacially deepened canyon of the Merced River and is located about halfway up the western slope of the Sierra Nevada, in middle California. It is about 3,000 feet in depth and steep-walled. Stretching northward and southward from the valley rims are rolling uplands, parts of an old land surface developed before the recent uplift and trenching of the range.

The rocks in which Yosemite Valley is carved are coarse crystalline intrusives ranging in composition from granite to gabbro. As shown on Calkin's geologic map[25] the three-dimensional pattern is quite complex, due to a succession of intrusions, each cutting the earlier rocks. Glaciation has removed the soil and mantle of decayed rock and thereby facilitated geophysical measurements.

Yosemite Valley was selected for experimental purposes both because

[25] "United States Geological Survey Professional Papers," No. 160, Plate 51, 1930.

of the types and favorable exposures of the rocks and because it appeared that it might be possible to determine whether a deep canyon or trench would eliminate the surface waves in a record made on one side of the trench by an explosion on the other.

Three localities were utilized for observations. All three are in the area mapped by Calkins as El Capitan granite. It is a medium to coarse-grained rock, contains quartz and biotite, and is streaked on the canyon walls by large and small inclusions of darker granitic rocks ranging from diorite to gabbro in composition. While some of the bodies of intrusive rock in the Yosemite region are markedly devoid of joints, the areas in which the experimental work was carried on are jointed to about average degree.

Petrographic analyses of the granites.—Thin sections of the granite at both Tamarack Flat and Cascade Creek were studied microscopically by Dr. Ian Campbell, professor of Petrology in the California Institute of Technology, to whom the thanks of the authors are due. Dr. Campbell's report is as follows:

Under the microscope the specimen from Tamarack Flat is seen to be a rather coarse-grained, equigranular, soda-potash granite. The chief constituents, together with their percentage distribution, are:

	Percentage
Quartz	41
Oligoclase ($ab_{85}an_{15}$)	37
Microcline	16

These percentages represent the average of Rosiwal analyses made on three sections, each cut from the same hand specimen, and averaging four square centimeters in cross-sectional area.

Biotite (6 per cent) is a fairly abundant accessory; apatite, muscovite, titanite, and zircon are present in scant amounts. The oligoclase shows a slight zoning, from about ab_{83} at the center to ab_{86} on the outer edge of the crystal.

The average grain size is from 1 5 to 2 centimeters The crystallization of the constituents has followed the usual order. The rock is very fresh: the only alteration being a slight chloritization of the biotite and some incipient epidotization of the oligoclase. The quartz shows a small amount of fracturing, while strain shadows are practically absent.

In thin section the specimen from Cascade Creek appears as a medium-grained, slightly porphyritic, potash-soda granite The principal constituents, together with their percentage distribution, are:

	Percentage
Oligoclase ($ab_{72}an_{28}$)	31
Orthoclase	29
Quartz	27
Biotite	8
Microcline	5

These percentages represent the average of Rosiwal analyses made on two sections, each cut from the same hand specimen, and averaging five square centimeters in cross-sectional area

Titanite is the most abundant of the accessory minerals, but is present in amounts less than 1 per cent. Apatite, green hornblende, and zircon are present only in minute quantities

The average grain size is around two to three millimeters, but occasional crystals of quartz and of oligoclase measure up to one centimeter diameter. Crystallization has followed the normal order: microcline is distinctly later than the other constituents and might be interpreted as representing a deuteric stage. The oligoclase, particularly in the larger crystals, shows a certain degree of zoning, from about ab_{70} at the core to ab_{76} on the outside of the crystal.

There has been some chloritization of the biotite. All of the feldspars occasionally show a slight degree of kaolinization, and locally small amounts of epidote have formed at the expense of the more calcic portions of the oligoclase crystals. The quartz often exhibits a faint undulatory extinction On the whole, however, there has not been much alteration and only negligible strain in this rock

Wave-velocity in granite and investigations on the thickness of the granite layer.—Investigations using the records of earthquake waves have shown that beneath the sedimentary layers there is a layer with a velocity of 5½ to 5¾ kilometers per second for longitudinal waves and of 3¼ kilometers per second for transversal waves. The maximum thickness found up to date for this layer appears to be beneath the Alps, where a thickness of over thirty kilometers seems to occur; on the other hand this layer seems to be absent in central and northern Germany, where the velocity of the longitudinal waves in the uppermost rocky shell is 6.0 ± 0.2 kilometers per second. Investigations of earthquakes and blasts in California up to date have always shown[26] that the velocity of longitudinal waves in this layer is between 5.5 and 5.6 kilometers per second with a thickness of the order of fourteen kilometers

It seems very probable that this layer consists of granite, but the velocities of elastic waves in granite, which have been calculated from the elastic constants found in laboratories, especially at the Geophysical Laboratory of the Carnegie Institution of Washington, range between 4¾ and 5½ kilometers per second. In these experiments a pressure corresponding to that at a depth of seven kilometers has been used, which means that their average is noticeably less than the values ob-

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[26] P. Byerly, *Bulletin of the National Research Council*, No. 61, p 88, Washington, July 1927; H O. Wood and C F. Richter, *Bulletin of the Seismological Society of America*, **21**, 28, 1931, and **21**, 183, 1931; B. Gutenberg, *Gerlands Beiträge zur Geophysik*, **35**, 6, 1932

served in the "granitic layer" of the earth. R. A. Daly[27] tried to explain this difference by the hypothesis that it may be caused by a different effect of the small stresses occurring in earthquake waves and the large stresses applied in the laboratories. For these reasons it would be very helpful in many geophysical problems to find the thickness of this layer and the wave-velocities in granite by use of the records of artificial explosions made primarily for this purpose.

A region especially suitable for this purpose is Yosemite National Park. Wide areas of this Park consist of granites which, owing to glaciation, are fresh at the surface or are covered only by a very thin alluvial layer. Moreover, the facilities of the Park (good roads to many points, and telephone lines which could be used through the courtesy of the Park Service) aid greatly in field operations. However, the principal reason for choosing the Park for this investigation was the idea that the steep walls on both sides of the Valley (for example between Glacier Point and Yosemite Point) would facilitate the determination of the thickness of the granitic surface layer. If the explosion takes place on one rim and the instruments are installed on the other rim, direct waves and surface waves will not travel down one wall, pass through the valley, and pass up the other wall. Therefore only diffracted and reflected waves should be registered.

Upon arrival in the Valley it was found that the tunnel for the new all-year Wawona Road was under construction. It is located between the old Wawona Road and the Valley, north of Inspiration Point. Blasts were being fired in groups about twice daily, and the waves produced by them could fortunately be utilized. Customarily the blasts of each group were timed to follow one another within a few seconds. The contractors agreed to increase the time-interval between the first and the second explosion, up to nine seconds in some cases, so that reflected waves arriving within that interval could be registered without being disturbed by the second shot. Supposing a velocity of five kilometers per second, reflected waves from depths down to at least twenty kilometers could have been registered before the first waves of the second explosion arrived.

Unfortunately the slope of the sides of the Valley is considerably less in this locality than it is farther east, the slope from Gentry Checking Station, where the instruments were installed for the first four

[27] R. A. Daly, *American Journal of Science,* **15**, 122, 1928; *Gerlands Beiträge zur Geophysik,* **19**, 198, 1928, and **22**, 26, 1929.

explosions, to the bottom of the Valley being less than 30°. In all cases the seismograms begin with small waves, followed by irregular larger waves during the first one-half-second, then the amplitudes decrease gradually. No reflected waves could be found. Therefore an attempt was made to register the waves closer to the point of explosion, as in this case the reflected energy should be larger, as we have seen on p. 207. One blast was registered on the new road, one mile from the mouth of the tunnel, but no reflected waves were recorded. The last shot was registered on the old Wawona Road. In this case about 500 pounds of dynamite were fired at the first shot. One instrument was nearly vertically over the point of explosion, at a distance of about 200 meters. The beginning was registered 0.04 of a second after the explosion. But, in this most favorable case also, no noticeable reflected waves can be found within the time interval between the first and the second shock —over eight seconds—so that either even this energy was not large enough, or the thickness of the granitic layer is more than twenty kilometers. As there was no hope of getting the reflected waves with the energy available in our investigations, the experiments on this problem were discontinued and no attempt was made to study waves which had passed under the steeper walls of the eastern part of the Valley.

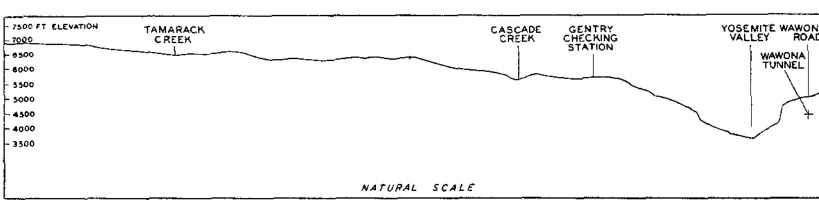

FIG. 7—Surface profile across lower Yosemite Valley, from new Wawona Road tunnel to Tamarack Creek, indicating slope of valley walls, relation of shot-point in tunnel to recording points on Wawona Road, and at Gentry Checking Station, and geographic location of Cascade Creek and Tamarack Creek, where velocities in granite were measured.

In some other cases, especially during the following investigations on the velocity of the waves in granite, charges up to fifty pounds were fired under conditions in which a very large fraction of the energy seemed to have entered the granite, but no waves were found which could have been reflected from the lower boundary of the granite. On the other hand, only at distances greater than 100 kilometers may waves refracted at this surface be registered at the beginning of seismograms, so that their investigation by use of blasts is restricted to occasional industrial blasts with very large quantities of dynamite; therefore it is very desirable that the preparation for such blasts be communicated very early to seismic stations in the region where the blast is to occur, so that plans for recording of the waves and accurate timing of the blast itself may be made.

The other problem considered was the determination of the velocity of the longitudinal waves in granitic rock. In this case it is desirable to have a large, plane area of unweathered rock at the surface, at a place which can be reached with instruments. All these conditions are very seldom fulfilled at the same time. Two suitable places were found. The first was Cascade Creek, north of Gentry Checking Station, where the profile ran along the creek; and the other was near Tamarack Flat, about two and one-half kilometers (one and one-half miles) northwest of Gentry's, where the instruments were installed west of the Big Oak Flat Road and parallel to it. Explosions were set off at several points to the southeast along a line nearly parallel to the road. At Cascade Creek all the instruments were set directly upon the granitic rock. The charges were put into small holes in the granite, filled with water. The fraction of energy which went into the rock must have been considerable, since charges amounting to only a small fraction of a pound of dynamite gave very sharp beginnings on the seismograms. The distance in this case was limited by the configuration of the creek. Three different shot-points were selected with twelve shots altogether (for elastic waves and sound waves separately), which were fired within an interval of four hours. The observed travel-times are given in Table VII, together with the travel-times observed near Tamarack Flat. There a larger area could be used, but a thin layer of alluvium covered the granite. Instruments and shot-points were in the alluvium. At many points hereabouts granitic rocks protrude through the alluvium, reaching to a few meters above the surface. In this case, too, three shot-points were used with eight shots altogether, fired within an interval of four hours.

The following straight lines seem to fit the data best, considering that the sum of the squares is least in these cases and that no large differences occur near the beginning or the end of the profiles:

$$\left.\begin{array}{ll} \text{Cascade Creek.} & t = 0{\cdot}002 + \dfrac{\Delta}{5.25} \\[2ex] \text{Near Tamarack Flat.} & t = 0{\cdot}025 + \dfrac{\Delta}{5.25} \end{array}\right\} \Delta = \text{distance in kilometers}$$

TABLE VII

TRAVEL-TIMES OF LONGITUDINAL WAVES* AND DIFFERENCES OBSERVED MINUS CALCULATED TRAVEL-TIMES

Cascade Creek			Near Tamarack Flat		
Distance	Time	Difference	Distance	Time	Difference
53	13	1	209	52	(—15)
61	14	0	238	70	0
107	21	—1	332	90	2
165	33	0	380	101	4
210	45	3	494	117	—2
163	34	1	412	99	—4
184	37	0	441	110	1
225	41	—4	522	128	3
270	52	—1	565	138	5
298	60	1	670	156	3
273	55	1	644	141	—7
296	57	—1	681	155	0
322	62	—1	747	163	—4
384	73	—1	784	174	0
412	84	3	831	184	1
			944	205	1

* Distance in meters, time in thousandths of a second.

The travel-time is plotted against the distance in Figure 8. The values relating to Cascade Creek fit very well; the values in the other case show larger differences. The first point very probably belongs to

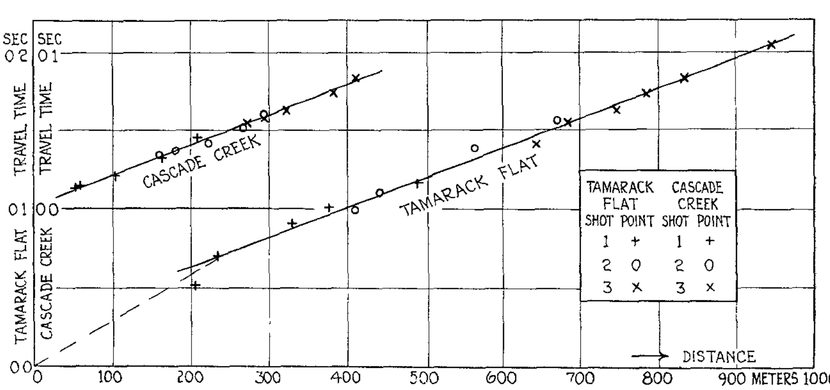

FIG. 8.—Chart showing travel-times through granodiorite (Yosemite Region)

the travel-time curve through the upper layer. In addition, the first instrument in all cases shows earlier arrival of the waves than corresponds to the average. Probably it was closer to the granite. On the other hand, the waves from the second shot-point arrived a very few thousandths of a second too late. Perhaps near this shot-point the surface of the granite was somewhat deeper than under the other two shot-points. Considering all sources of errors it seems very probable that the velocity of the longitudinal waves in the granite of the regions where the explosions took place is between 5.1 and 5.4 kilometers per second, and that $5\frac{1}{4}$ kilometers per second is the best value available from our data.

There is no doubt that the velocity of 5.5 to 5.6 kilometers, calculated from blasts and earthquakes, is higher than the velocity in the granite of the region where our investigations were made. As the velocity of the waves in the "granitic layer" nearly everywhere seems to be higher—in the Alps only velocities of 5.4 kilometers per second or even less have been observed—it appears very reasonable that the material of the "granitic layer" is more rigid and less compressible than the granite considered above. These differences, and differences between the elastic constants found from elastic earth waves and from laboratory tests, may be caused by pressure, by local differences of mineral composition which affect the results derived from samples while seismic methods give average values, or finally by the differences in velocity in different directions in gneissic or schistose structures. Still lower velocities were found in surface granite by L. Don Leet and Maurice Ewing:[28] 4.96 ± 0.02 (probable error) kilometers per second in Quincy granite (where the velocity of transversal waves was 2.48 ± 0.03 kilometers per second): 5.00 ± 0.04 kilometers per second in Westerly granite; and 5.08 ± 0.01 kilometers per second in Rockport granite.

Thickness of Sediments in Los Angeles Basin

Geology.—The Los Angeles plain is a smooth land surface sloping gently southwestward, interrupted here and there by low hills, and extending for distances up to a few tens of miles west, south, and east of that city. While locally discernible, the structure of the folded formations underlying this plain is obscured over large areas by alluvial deposits spread over it by streams crossing it from the mountains at the north.

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[28] L. Don Leet and Maurice Ewing, "Velocity of Elastic Waves in Granite," *The Physical Review*, second series, **39**, 868, 1932, *Physics*, **2**, 160, 1932

The zone along which geophysical experimenting was done is southeast of Los Angeles and extends from a point northwest of Wilmington (3,000 feet west of Harbor Boulevard and 1,000 feet south of Santa Fe track) to a point two miles east of Norwalk, a distance of about seventeen miles. This area was chosen because of geological considerations and because it is less densely settled than most other parts of the basin.

Because of the alluvial mantle and the fact that wells have penetrated only the upper part of the thick sedimentary section, the geology of the district is not fully known.

The northwest-southeast Inglewood fault crosses the belt of shotpoints about five miles northeast of the southwest end. West of the fault relatively soft Miocene strata rest on somewhat more compact Jurassic Franciscan sediments. Presumably granite or highly crystalline metamorphics underlie the latter.

East of the fault the plain is underlain by a thick section of relatively soft sandstones and shales, largely marine in origin and mainly of Tertiary age. It is possible that somewhat more compact Cretaceous sediments or Jurassic Franciscan strata underlie the Tertiary formation. If so the change in physical properties from the upper relatively uncompacted part of the section to the crystalline rocks below is much less abrupt than if the middle Tertiary strata rest directly upon the crystallines. The total thickness of the sedimentary section is quite certainly over 20,000 feet (6,000 meters) and it has been suggested by some geologists that it may attain 40,000 feet.

The structure of the area studied is known to be synclinal, with an axial trend of northwest–southeast. The sides of the fold dip gently. Bodies of sandstone of considerable thickness alternate with the less rigid shale strata. The gentle slope of the beds and contrast in physical properties of the shale and the sandstone members led to the expectation that conditions for wave reflection would be favorable.

Seismic investigations.—The first problem attacked was to find the thicknesses of the sedimentary layers in Los Angeles Basin and especially the depth at which the granite lies. Similar investigations had previously been made in San Joaquin Valley, where for example F. E. Vaughan,[29] by using the refraction method, had found that the granite

[29] F. E Vaughan, "Some Applications of Geophysical Methods to Theoretical Studies in Geology" Paper presented at the meeting of the American Association for the Advancement of Science, Pasadena, June 19, 1931.

of the Sierra and the Coast ranges is continuous beneath the sediments of the valley, and that the thickness of the sediments is, for example, 800 meters at a point thirteen kilometers northwest of Visalia, and three kilometers at a point six kilometers west of Hanford. In the Los Angeles Basin very much greater depths were expected and it therefore seemed very improbable that a profile of the length necessary for waves refracted through the granite could be found in this thickly settled area, so the reflection method was tried first. Fourteen shot-points were selected along a profile running from a point about one mile north of Wilmington to a point about one-half mile northwest of La Mirada (about three miles southeast of Norwalk). The different shot-points are shown in Figure 9.

As the ground was very dry at this time, it was often difficult to find favorable places in which to detonate the dynamite. To keep costs down, all holes were dug by hand. Usually one whole day was needed for every shot-point. The number of shots fired at one point varied between two and thirteen. The maximum charges were 10 pounds of dynamite, or less, at seven points; 10 to 20 pounds at six points; 50 pounds at two points, and 60 pounds at one point. Larger charges could not be used on account of the shallow holes and the dense population of the area.

At a few points the velocity in the upper layers was determined. It was found to be the same at all places within the limits of error We will consider it in connection with the refraction profile. At some points no reflections whatever could be found on the seismograms, while at other points the reflections were very clear, similar to the reflections reproduced in Figure 12 (Ventura Basin). In Table VIII, on page 236, the data for the reflections and the depths of the reflecting surface are given, for the most part, according to calculations by Mr. Salvatori. The data are not sufficient to permit following the different surfaces along the profile nor to determine whether any of the calculated depths correspond to the boundary between sediments and the granite. Consequently an attempt was made to get more information by shooting for refracted waves along a profile as long as possible A line containing the shot-points 7, 14, and 15 (Figure 9) was chosen. At the last shot at point 15 the instruments were placed as close to Bellflower, and south of it, as it could be arranged without encountering too much disturbance from traffic. Thus a maximum distance between shot-point and instruments of more than eight kilometers was utilized but, as had been expected, even this was too small to yield refracted

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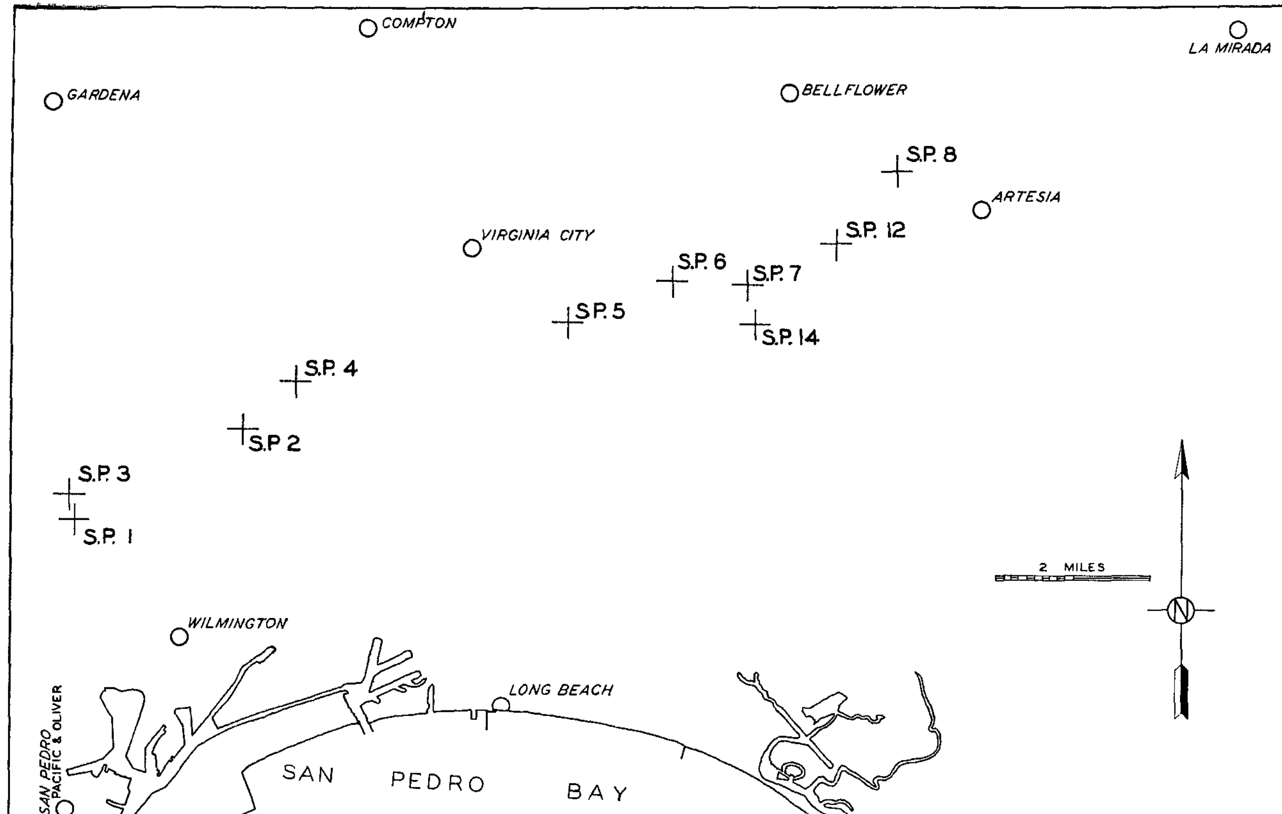

FIG. 9.—Sketch map of part of Los Angeles Basin indicating locations of shot-points.

TABLE VIII

REFLECTIONS ALONG A PROFILE IN LOS ANGELES BASIN†

S.P.	Distance	Time	Depths
1	368–613	1,110* 1,312*	1.13, 1.41
3	89–212	990	1.00
4	54–203	1,342*, *1,466*, 1,658, 1,709*, 1,909, 2,021, 2,520*, 2,591*, 2,682*	1.50, 1.64, 1.91, 2.0, 2.3, 2.5, 3.4, 3.5, 3.7
6	63–212	1,397, 1,466*, 2,452, 2,499*, 2,876*, 3,037*	1.54, 1.64, 3.3, 3.4, 3.9, 4.2
7	61–213	*1,197*, 1,468	1.27, 1.64
14	61–138	983*, 2,279*	1.00, 2.9
12	61–213	1,503*, 1,939, *2,053*, 2,123*, 2,214, 2,291*, *2,524*, 2,605*, 2,776*	1.68, 2.3, 2.5, 2.6, 2.8, 2.9, 3.4, 3.5, 3.9
8	63–212	*859*, 1,394, 1,516*, 1,704, 1,826*	0.85, 1.54, 1.70, 1.97, 2.16

† S.P = number of shot-point (Figure 9), Distance = distance between shot-point and the nearest and most remote instrument in meters; Time = time differences between the time of explosion and the arrival of the reflected waves in thousandths of a second; Depths = calculated depths of the reflecting surfaces in kilometers. Good reflections are italicized, fair reflections are marked by *; the other reflections are poor.

waves through the granite. The charges for the last two shots were forty and thirty-five pounds of dynamite. In all cases the instruments and the blasting machine were connected in the usual way and the telephone was used between shot-point and instruments. But at these long distances delays occurred because the connecting wires were broken several times and it required two days to secure the records for the five more distant groups of observations. The data calculated from these are given in Table IX together with data concerning certain smaller distances.

The Figure 10 travel-times along the profile 15–14–7 are plotted against the distance, together with some travel-times at shorter distances obtained from other shot-points. The travel-time Curve A shows clearly that there are several layers of sediments in which the velocity of longitudinal waves increases slightly from layer to layer with depth. If we use the Formulas 18 to 22, we find the following velocities of longitudinal waves in the different layers:

Depth in meters.	0–20±	20–320	320–860	860–1,650	1,650–?
Velocity kilometers per second	1±	1.9	2.1	2.9	3.5

When calculating these results, it was assumed that along a north–south line the surfaces are horizontal, which is not wholly true. Therefore,

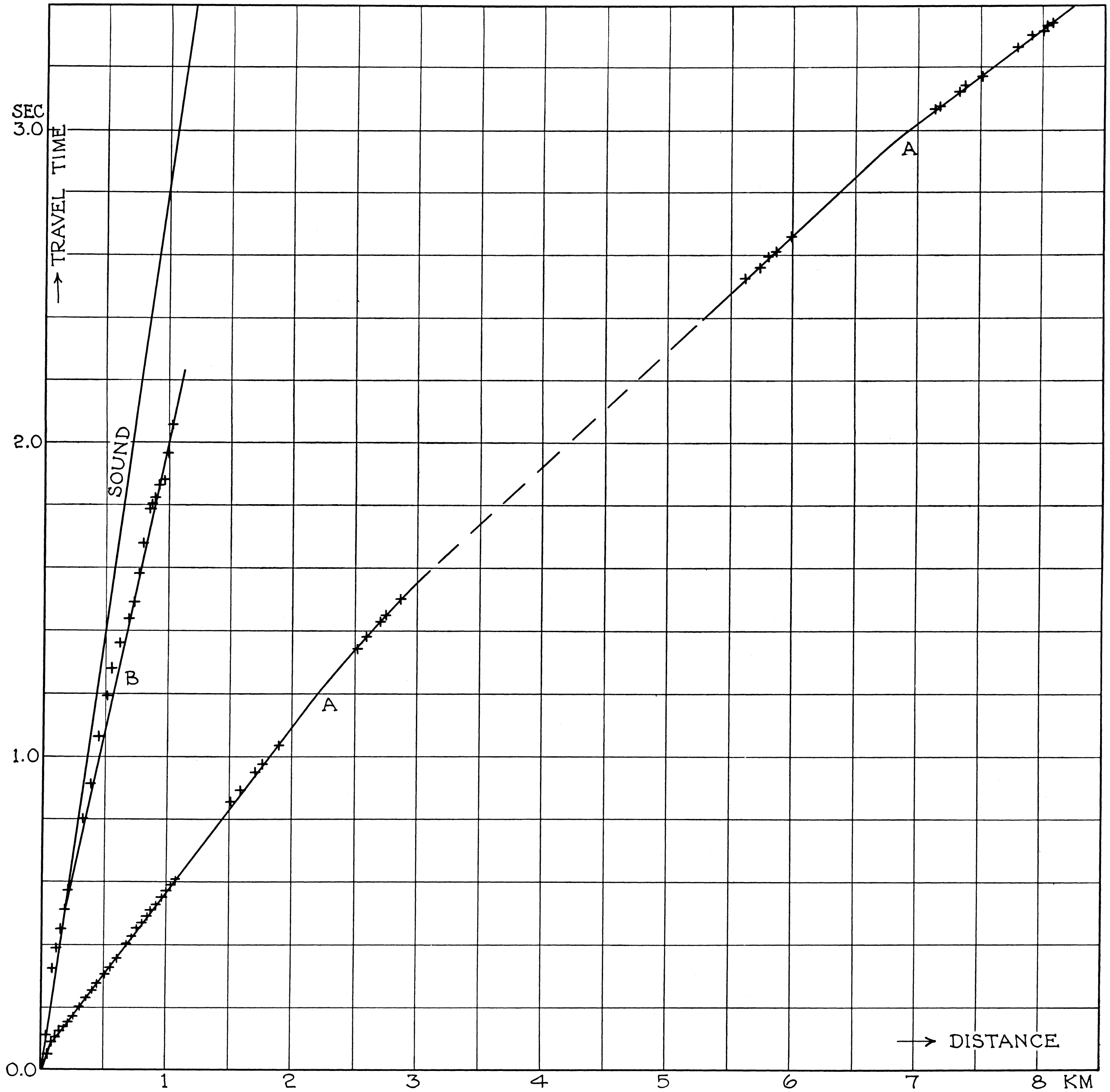

FIG. 10.—Chart showing travel-times in the Los Angeles Basin.

TABLE IX

Travel-Times of Longitudinal Waves in the Los Angeles Basin*

S P 1 North		S.P 2 West		S.P. 3 South		S P. 4 East		S P 14 North		S.P. 15 North	
Distance	Time	Distance	Time	Distance	Time	Distance	Time	Distance	Time	Distance	Time
61	54	188	129	89	87	54	62	61	50	5,629	2,528
92	81	235	155	119	102	84	80	90	69	5,746	2,562
122	97	261	171	150	122	113	99	122	88	5,802	2,597
152	115	311	200	182	137	144	118	153	102	5,861	2,615
183	129	352	225	196	145	173	135	184	120	5,990	2,659
214	145	402	250	212	152	203	153	214	139		
368	232	328	210	876	507			1,520	854	7,151	3.071
415	253	390	250	926	527			1,592	892	7,177	3,076
452	275	451	284	966	548			1,704	947	7,333	3,125
509	304	507	315	1,000	570			1,764	976	7,388	3,142
554	323	552	340	1,037	588			1,900	1,034	7,519	3.174
613	355	611	374	1,073	607						
						S.P. 5 South					
						Distance	Time				
		686	409			62	60	2,520	1.342	7,801	3,268
		723	425			92	75	2,596	1,383	7,918	3,308
		760	450			123	100	2,705	1,430	8,007	3,318
		802	467			153	120	2,750	1,452	8.039	3,327
		849	493			183	135	2,875	1,520	8,085	3,343
		897	517			214	152				

* S P = number of shot-point; the directions given mean the direction from the shot-point toward the instruments Distance = distance from shot-point to instrument in meters Time = travel-time in thousandths of a second

the values are somewhat uncertain From the data it follows also that the surface of the granite must be deeper than two kilometers. Comparing them with the results gained by the reflection method (Table VIII), we see that the discontinuity at a depth of 1 65 kilometers, where the velocity increases from 2.9 to 3½ kilometers per second, was found at the points 4, 6, 7, and 12, and that the reflections caused by it are in general fair or good The surface found by the refraction method at a depth of 860 meters is perhaps the same as the surface near one kilometer, found from reflections. As it is somewhat difficult to determine exactly the distance at which two parts of the travel-time curves with slightly different slope intersect, as can be seen from Figure 10, the reflection method is more accurate in finding the depth of the discontinuity in such a case; but the velocities in the two layers can be found very much better from the refraction method, as the interval in distances at which the reflections are observed is in general too small to yield good results by using Formula 10.

Returning now to Table VIII, we see that it is very improbable that any discontinuity found by reflections at a depth of less than three kilometers can be the surface of the granite, as depths between two and three kilometers occur in a few cases only. At a depth of about three and one-half kilometers there seems to be a well-marked discontinuity in the central part of the profile, but there is no other reason to think that this is the surface of the granite. From geological considerations it is probable that it is at a very much greater depth.

Besides the longitudinal waves and the sound waves, which latter were recognizable only when the dynamite was exploded at the surface or if the dynamite blew out the hole, in a few cases long waves with periods of one-twentieth to one-fortieth of a second have been observed, especially when the dynamic magnification of the instruments for such periods has not been artificially reduced contrary to the usual procedure; in general it is preferable to avoid recording these long waves, since they may disturb the reflected waves, which usually have short periods. Travel-times of such long waves have been plotted in Figure 10. Their apparent velocity has been found to be 550 meters per second in the Los Angeles Basin, and their wave-length there is of the order of twenty meters. At short distances they arrive somewhat later than the sound waves, at long distances earlier, but they are not related to them, as they are noticeable only in case the dynamite is buried, which is contrary to the case for the sound waves. Similar results have been found by Angenheister[30] in Juterbog, Germany, at a place where the alluvium has a thickness of about 100 meters.

Structure in Southeastern Part of Ventura Basin

Geology.—Experiments were carried out for a limited time in the southeastern part of the Ventura Basin, southeast of Oxnard, and about sixty miles west of Los Angeles. The instrument and shot-points lie in a zone extending northwestward from the base of the western part of the Santa Monica Mountains, which bound the Ventura Basin on the southeast.

[30] G. Angenheister, "Beobachtungen bei Sprengungen," *Zeitschrift für Geophysik*, 3, 28, 1927.

The mountains consist mainly of lower and middle Miocene sedimentary formations intruded by sills and dikes of diorite and basalt and overlain by upper Miocene extrusive basaltic and andesitic lava flows and agglomerates. The strata are folded and in general strike northeast and southwest. Southeast of Oxnard, in the environs of Round Mountain, the formations dip approximately 30° northwest. The volcanics are overlain, beneath the plain, by alluvium and by soft unconsolidated sand, gravel, and shale of upper Pliocene and Pleistocene age.

The Texas Eastwood well passed through these unconsolidated materials for 1,915 feet (584 meters), then entered the volcanics and after encountering several vesicular zones ended at 2,665 feet (813 meters) without reaching the base of the effusives. The well of the Mugu Syndicate, Incorporated, Ltd., Utt No. 1, also passed through soft materials to the volcanics at 1,600 feet (488 meters), passed through the volcanics at about 5,575 feet (1,700 meters), and ended in Temblor sediments at 5,866 feet (1,790 meters).

There is reason to think that two northeast-southwest faults, one passing just west of Leesdale, the other just west of Elrio, traverse the area, and that a third, trending northwest-southeast, passes through Sucrosa.

The post-volcanic sediments are of course much less dense and rigid than the effusives on which they lie and the boundary surfaces between them would, it appeared, reflect waves effectively.

The district is therefore one in which the geology is concealed but in which the stratigraphy and structure are nevertheless known in a general way and it seemed therefore a favorable locality for a geophysical test.

Seismic investigations.—As it became evident from the investigations in the Los Angeles Basin that no discontinuities at depths of more than a few kilometers could be found with the energy available in our investigations under normal conditions, the region between the hills which form the southern boundary of the Ventura Basin and the region around Oxnard were chosen for another attempt to find the thickness of sediments. According to the geological description given above it was to be expected that the basalt forming the hills at the south continues under the sediments with a certain slope to the north, that beneath the basalt there are perhaps other sediments, and that the granite surface is at greater depth. For this investigation twenty-two shot-points were used. Their positions are marked in Figure 11 (p. 240). In general eight to ten shots were fired at each point. At six points only one or two shots were

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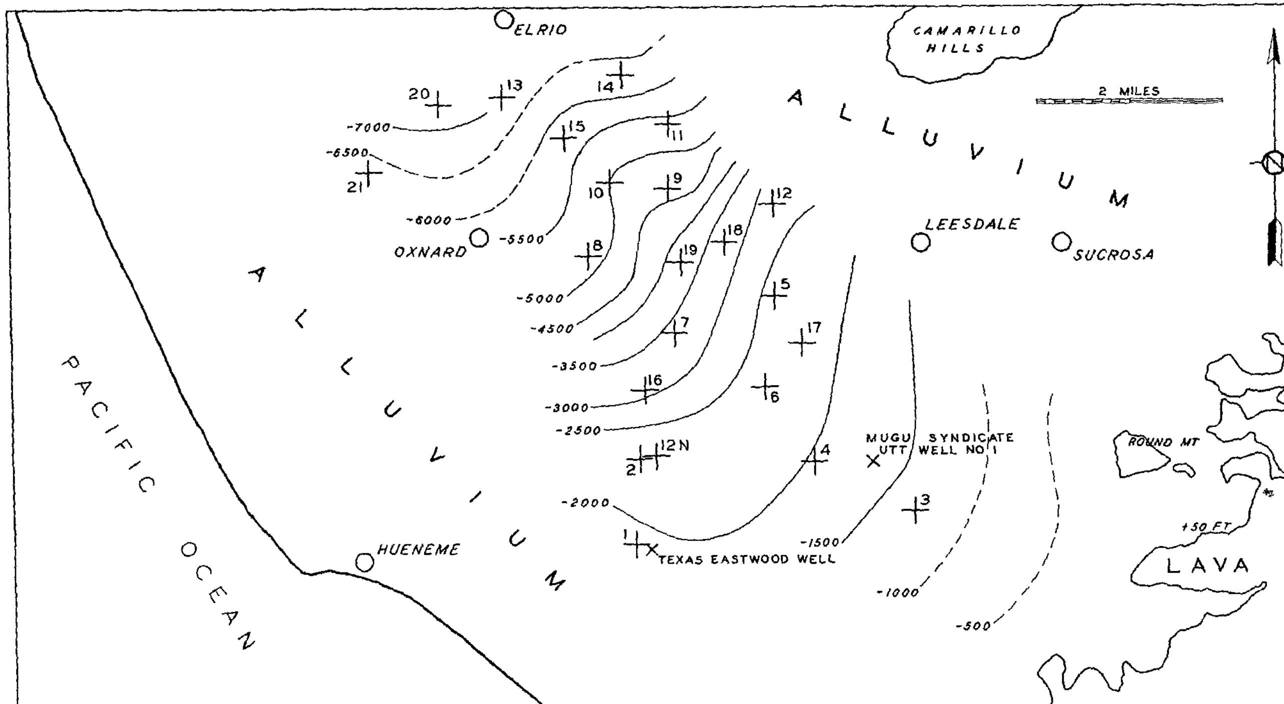

FIG 11.—Sketch map of part of Ventura Basin, indicating relation of alluvial plain to lava foothills of Santa Monica Mountains on the southeast, and location of numbered shot-points and two wells Lava dips northwestward, as is indicated by structure-contours drawn on its upper surface, on basis of seismic measurements (structure-contour interval, 500 feet). As indicated by map, lava was traced from the surface to a depth of over 7,000 feet

fired with charges of more than ten pounds of dynamite, with a maximum of fourteen and one-half pounds at point 17. All the shooting was done between July 27 and August 5, 1931. All calculations concerning depths were made by Mr. Salvatori, who also directed the field work.

The velocity of longitudinal waves in the upper layers was found to be nearly the same as in the Los Angeles Basin. In Table X a few values for travel-times are given, together with corresponding values from explosions in the Los Angeles Basin. At short distances (less than 100 meters) the waves arrived somewhat later in the Los Angeles Basin, at longer distances (some hundred meters) somewhat earlier there than in the Ventura Basin, but the differences seldom exceed one-hundredth of a second. The average velocity in the upper layers therefore is nearly the same in both regions. No attempt was made to find the velocities at greater depths in the Ventura Basin In the following calculations Mr. Salvatori assumed that the average velocity at a given depth is the same in both basins.

TABLE X

TRAVEL-TIMES OF LONGITUDINAL WAVES†

L.A. 2		V. 2 West		V. 7		V. 1		L.A. 1	
Distance	Time	Distance	Time	Distance	Time	Distance	Time	Distance	Time
255	155			244	161	46	35		
261	171			274	180	61	44	61	54
311	200			305	197	92	65	92	81
352	225	352	230	366	236	122	84	122	97
390	250	382	250	396	255	152	109	152	115
451	284	441	287			183	129	183	129
507	315	472	306			214	148	214	145
						228	159		

† Units as in the previous tables L.A. = Los Angeles Basin. V = Ventura Basin The number following is the number of the shot-point.

As the velocity of the longitudinal waves was found to be nearly the same, it was expected that the "long waves" also would arrive after the same time interval in both regions, but this did not occur. The "long waves" arrived very much later in the Ventura Basin, under similar experimental conditions, than they did in the Los Angeles Basin. A few data are given in Table XI. Their velocity generally was somewhat less than the velocity of sound, and their travel-times always are greater than the travel-times of the sound waves. In general their period was of the order of 0.035 of a second, so their wave-length is of the order of ten meters only. It is very difficult to present any hypothesis as to the nature of these waves without more data registered simultaneously by a set of instruments with three components. The most plausible hypothesis is that they are Rayleigh waves in the uppermost layer, but in this case neither their late beginning near the point of explosion nor the different velocities in the two regions considered above is understandable. As on the other hand their velocity is of the order of one-half kilometer per second in the Los Angeles Basin and of the order of one-third kilometer per second in Ventura Basin, the corresponding velocity of longitudinal waves should be less than one kilometer per second in both cases, and this is the case very near the surface only.

TABLE XI

TRAVEL-TIMES OF LONG WAVES AND OF SOUND WAVES IN VENTURA BASIN*

Shot-Point 2		Shot-Point 7		
Long Waves		Long Waves		Sound Waves
Distance	Time	Distance	Time	Time
150	761	137	680	393
180	860	305	1,180	874
200	980	396	1,460	1,134

* Distance in meters and time in thousandths of a second

Many clear records of reflected waves were registered from the explosions in the Ventura Basin and there were only a very few points where no reflected waves could be found. The reflections may be divided into three groups: one consists of a few reflections, occurring within an interval of about one-tenth of a second and showing a quite regular increase in travel-time from the southeast toward the northwest. Reflections of the second group arrive earlier than these, reflections of the third group later. As the depths of the reflecting discontinuities in the case of the first group, as calculated from the records at the points 1, 3, and 4, correspond very well to the depth of the basalt found in the two wells in this region, it seems very likely that all reflections of this group are caused by the several layers of basalt which are to be expected beneath the alluvium in the region bounded on the southeast by the basaltic hills and, on the northeast and northwest, by fault zones, as has been considered in detail in the preceding section. The depth of the

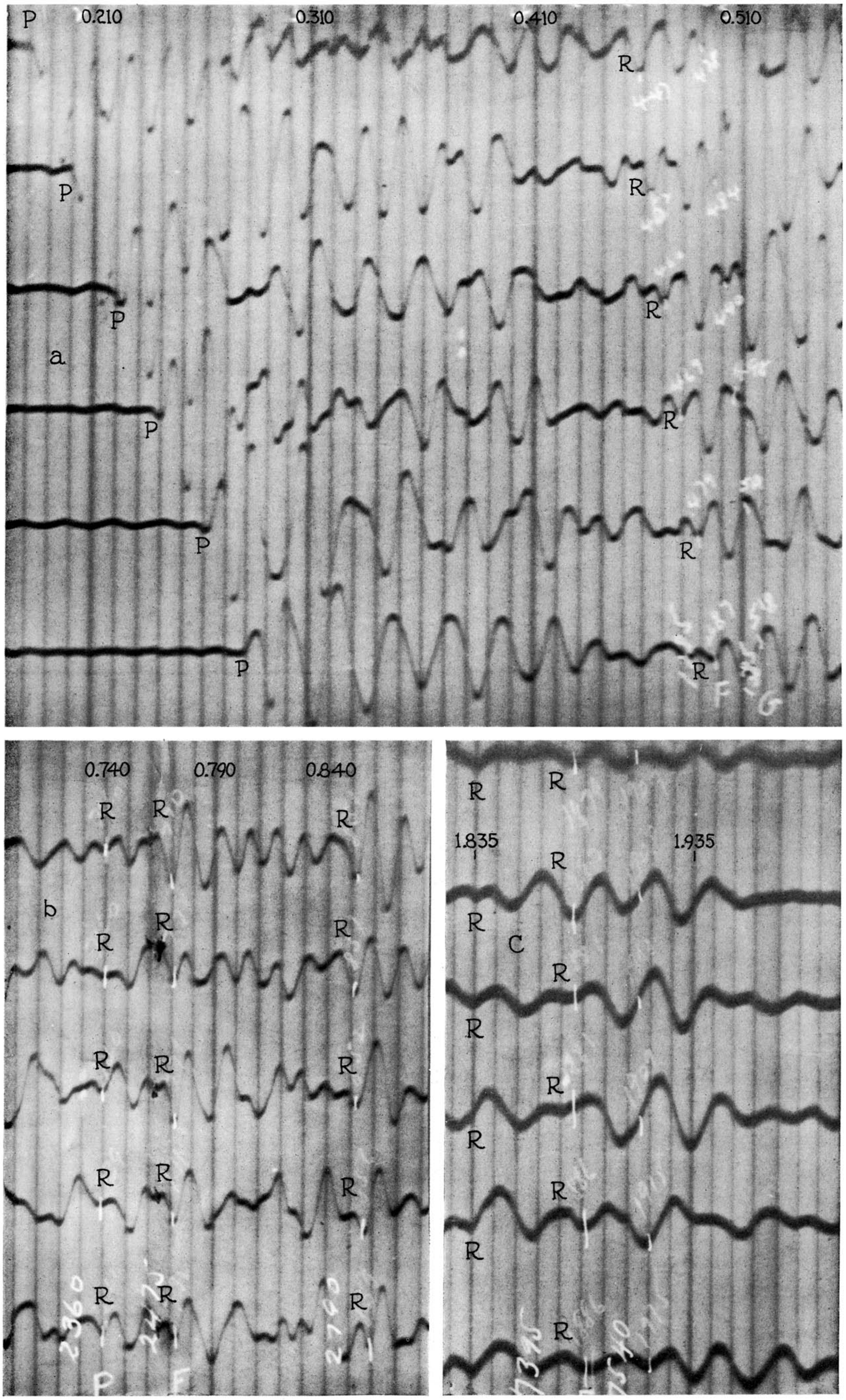

FIG. 12.—Seismograms from Ventura Basin: (*a*) at shot-point 3, distances between shot-point and instruments 275 to 426 meters, charge, ¼ pound of dynamite; (*b*) at shot-point 12, distances 60 to 122 meters, charge, 1½ pounds; (*c*) at shot-point 20, distances 46 to 122 meters, charge, 10 pounds of dynamite. The vertical lines give the time at intervals of 1/100 of a second. The figures at the top of the seismograms indicate the time from the moment of explosion.

reflecting surfaces of the basalt layers, as calculated from the seismograms, is given in Table XII. Three characteristic seismograms showing these reflections are reproduced in Figure 12.

TABLE XII

TRAVEL-TIMES OF REFLECTED WAVES* AND DEPTHS OF THE REFLECTING SURFACES OF BASALT IN THE VENTURA BASIN

Shot-Point	Travel-Times	Distance in Feet	Depth of Discontinuities	
			In Feet	In Meters
1	618	625	1.895	580
2 North	676	625	2.110	640
2 West	758	1,370	2,330	710
3	442. 469	625 and 1,150	1,285. 1.385	390, 420
4	645. 688. 735, 778	620	1,990, 2,155, 2,325, 2.480	610, 660, 710, 760
5	728	825	2,280	700
6	733, 804	650 and 1,050	2,315, 2.520	710, 770
7	Reflections poor	325	?	?
8	1,431. 1,460. 1,494	270	5,185, 5,320, 5,475	1.580, 1.620, 1,670
9	1.224, 1,322	270	4.275, 4,700	1,300, 1,430
10	1,388. 1,426	270	4.995, 5,165	1,520, 1,580
11	1,491, 1,528, 1.564	270	5,460, 5,630, 5,805	1,670, 1,720, 1,770
12	855, 908, 961	270	2.790, 2.990, 3,190	850, 910, 970
13	Sand prevents use of larger charges		?	?
14	1.650, 1.720	270	6.300, 6,600	1,920. 2,010
15	Reflections poor		?	?
16	951. 987	325	3,095, 3,235	940, 990
17	671	270	2,020	620
18	980. 1.057	250	3,270, 3,575	1,000. 1.090
20	1,882, 1.910	325	7,395. 7,540	2,260, 2.300
21	1,640, 1,706	360	6,250, 6,500	1,910, 1,980

* In thousandths of a second. Distance = average distance of instruments from shot-point in feet (1 foot = 0.305 meter)

Figure 12*a* shows the direct longitudinal waves and the reflected waves on a seismogram, recorded from shot-point 3 The charge was one-fourth of a pound of dynamite, the distances of the instruments from the shot-point between 900 and 1,400 feet (275 and 426 meters). One sees very clearly the difference between the direct waves, which arrive at the successive instruments at intervals of about 0.02 of a second, and the waves reflected from the basaltic layer, which arrive about one-half of a second after the explosion and whose times of arrival change less than 0.01 of a second from instrument to instrument. Between 0 3 and 0.4 of a second after the explosion there are other reflected waves. They are not clear on the record reproduced, but are on other records, taken at smaller distances from the shot-point. The depth of the basaltic layer near shot-point 3 is about 400 meters. Figure 12*b* is the reproduction of a seismogram from shot-point 12. The distances of the instruments from the shot-point were between 200 and 400 feet (61 to 122 meters). The charge was one and one-half pounds of dynamite. The depths of the reflecting layers are somewhat less than one kilometer. Finally, Figure 12*c* shows the waves reflected from the deepest point of the basalt surface found in this investigation, at a depth of about 2.3 kilometers. It was recorded at shot-point 20, in which the distance of the instruments from the shot-point was between 46 and 122 meters and the charge ten pounds of dynamite.

The lines of equal depth of basalt (structure contour lines on basaltic surface) are shown in Figure 11. The contour interval is 500 feet

As has been stated above, two additional groups of reflections were registered, one from shallower and one from deeper surfaces than the basalt. As may be seen from Table XIII showing their depths, the flat discontinuities were observed more especially in the eastern section of the region, and it seems that there are two different groups. The data available were not sufficient for defining structure contour lines.

Discontinuities deeper than the basalt (perhaps the lower boundary of the basalt in some cases) were found a few times as can be seen from Table XIII, also. A fair reflection was obtained from the explosion at shot-point 8, 3.187 seconds after the explosion was fired. As the velocities at larger depths are not known, the depth of the reflecting surface cannot be calculated accurately. It may be of the order of five kilometers. The depths given in Tables XII and XIII are not wholly correct, as the change of velocity with depth has been assumed to be the same as in the Los Angeles Basin The seismograms which show refracted waves demonstrate this discrepancy to a certain extent; at least from the surface of the basalt down the velocity is very different in the two regions. In the basaltic layer especially the velocity is higher than at the same depth in the Los Angeles Basin. Therefore the depths given in the last column of Table XIII probably are somewhat too small. It had been intended to run a refraction profile parallel to the contour lines of the basalt in the neighborhood of point 3 or 4 in order to get better information regarding these velocities, but this was not done on account of lack of time. When considering the data, it must always be remembered that the task set was not to secure final results, but to test methods of applied seismology in the investigation of geological problems.

TABLE XIII

DEPTHS OF REFLECTING SURFACES IN VENTURA BASIN

Shot-Point	Shallower than the Basalt		Deeper than the Basalt	
	Depth in Feet	Depth in Meters	Depth in Feet	Depth in Meters
2 North			2,450	750
2 West			2,685	820
3	930, 1,010	280, 310		
5	1,765	540	7,520	2,300
6			5,030, and more than 8,000	1,530 > 2,500
7	2,115	650		
10	3,700	1,130	6,600	2,000
11	4,600	1,400	6,400	1,950
12	1,700, 1,915, 2,360, 2,475	520, 580 720, 750		
14	2,500, 3,000	760, 920	$\geqq$10,000	$\geqq$3,000
16	1,855	570	5,265	1,600
17	1,710	520	4,500 > 8,300	1,370 > 2,500
18	1,865	570		

CONCLUSIONS

The purpose in performing the geophysical experiments recounted in this paper was not to secure certain seismic or geologic data, but primarily to test the effectiveness of seismic methods in the exploration of the earth's crust. It may be profitable therefore to summarize in general terms the conclusions reached. It seems to the authors that the experiments demonstrate that:

The velocities of surface waves and of direct compressional waves through the surface parts of the crust can be measured with a high degree of precision. Data secured in such measurements are of course of great interest and use in seismological research.

By the reflection method, depths to boundary surfaces between rock masses or strata possessing different physical properties can be measured down to 10,000 feet (3 kilometers) and, under the most favorable conditions, probably down to 20,000 feet or more. When reflections are recorded from successively lower boundary surfaces the thicknesses of the intervening rock bodies or beds are of course derived directly. Likewise folded structures can be identified accurately by securing the elevations of particular horizons or of the upper surfaces of certain beds at a considerable number of points. Sudden changes in depth of a particular reflection surface between strata or its disappearance laterally gives information regarding the presence and perhaps the amount of the vertical displacement of a fault. By reflections recorded from the fault surface a fault of low or moderate dip can be traced downward and its angle of dip determined.

By the refraction method velocities not merely in the superficial but also in the deeper layers can be accurately determined by spacing recording and detonation points relatively far apart. Depths to marked boundary surfaces can be computed. With sufficient depth data for a given discontinuity, structure of course becomes evident. The existence of a fault can be proved even if concealed, and the amount of its displacement inferred if conditions are favorable, but it is probably more difficult to determine its dip by the refraction than by the reflection method

The effectiveness of the seismic methods depends very much upon the favorableness of soil, formational, and structural conditions. Dryness or wetness of soil, depth to the water table, physical properties of the formations involved, the presence, condition, and form of structural surfaces or discontinuities, the nature of the folded or faulted structures under investigation, and apparently still other factors, some of which are rather obscure as to nature and effect, all seem to play important parts in determining the degree of success attainable by the methods in a given locality or region.

While the methods are obviously applicable in areas in which ordinary structural studies involving study of outcrops and/or well logs cannot succeed because of inadequate exposures or well data, it is also evident that the methods require rather elaborate equipment of highly specialized type, which cannot be purchased but must be built, and a highly trained personnel including individuals having expert knowledge of use of explosives, seismologic instruments, radio amplifying apparatus, interpretation of seismograms, structural geology, and the application of seismic data to structural problems. Explosives, photographic supplies, and other items tend to make the operations rather costly, as compared with ordinary structural studies in areas of abundant outcrops.

It is the conclusion of the authors that seismic methods constitute a powerful new tool for determining both the shallower and the deeper structures of the earth's crust.